



SOIL PHYSICS

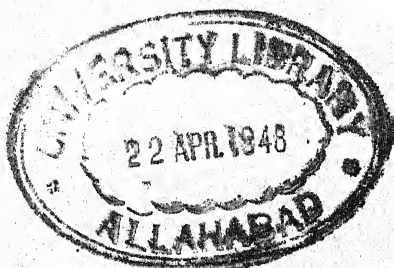


SOIL PHYSICS

BY

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GUY WOOLARD CONREY

MARION JACOB FUNCHES

MERRITT FINLEY MILLER

*in appreciation of their inspiration
and guidance*

PREFACE

Soil physics is a phase of soil science that has been receiving increasing interest and attention within the last twenty years. Numerous technical contributions have appeared in English, French, German and Russian scientific journals. In almost every instance, the individual has been interested in only one particular aspect of the physical properties of the soil. The field is so large that it limits the scope of activity of any one person.

Although much research in the field of soil physics has been accomplished, teachers of the subject have been handicapped by a lack of suitable instructional material. Unless the teacher happens to be closely associated with soil-physics research and has access to the numerous foreign publications, the preparation of a comprehensive course in soil physics is a difficult task.

The author has taught a course in this subject for the past nine years and has been fortunate in having had to review most of the foreign work in conjunction with his research projects. Practical experience in various aspects of soil-physics research, in addition to a rather wide coverage of French, German and Russian literature, has resulted in the preparation of a complete set of notes which have been used to build up a course in soil physics that has been rather favorably received by the students, especially those of graduate standing. Incidentally, the student reactions to the philosophy and content of the course have contributed much to the final shaping of the material into a form that can be easily and clearly presented.

An attempt has been made to discuss the various phases of soil physics from the point of view of the teacher explaining them to his class. Simple analogies and often extremely detailed discussions have been used to illustrate a significant point. The author has done this purposely, because it is his firm conviction that a subject cannot be taught successfully by assuming that the student is fully aware of the implications of the subject that seem so obvious to the highly specialized teacher. The book is designed primarily for the advanced undergraduate students of fairly good caliber and for graduates. There are several sections that undoubtedly will prove too involved for strictly undergraduate classes. However, they may be omitted

without materially affecting the continuity of thought in the book. Detailed mathematical discussions were purposely avoided for the sake of clarity. They may be added by the teacher who desires to use them.

The book does not propose to give complete citation of all published literature on a particular subject. The references at the end of each chapter have been selected because the author felt that they represented the more important points of view. Considerable reference has been made to some of the classical soil-physics research that was carried out during the last twenty-five years of the past century. This has been deemed essential, since many of these publications are not universally accessible.

In many instances, experimental evidence does not permit the formulation of an exact statement of fact concerning a particular property. The author has attempted to present the different viewpoints as he sees them and has stated his own personal opinions regarding these viewpoints. His opinions, of course, are subject to change as soon as further research data warrant.

It is realized that other investigators may view a given phenomenon differently from the way it is discussed in this book. Nevertheless, the material is presented to represent the author's concept of the subject. Differences in interpretation are usually only differences in points of view. The author has tried to develop a concept of the fundamental aspects of soil physics and their practical application. Special attention has been given to the practical interpretation of each of the different phases of the subject. It is believed that this philosophy will make the book not only more interesting but also more usable.

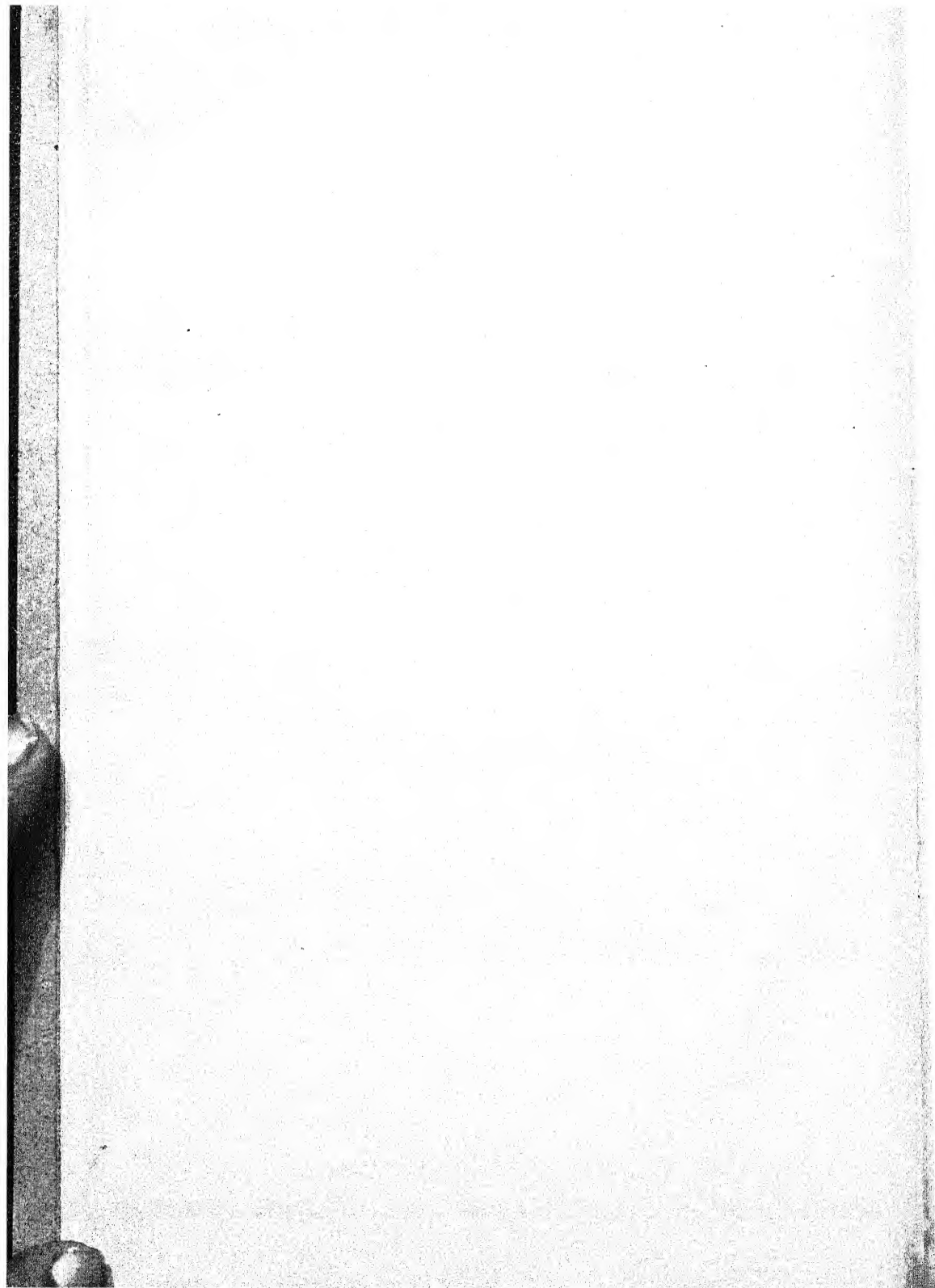
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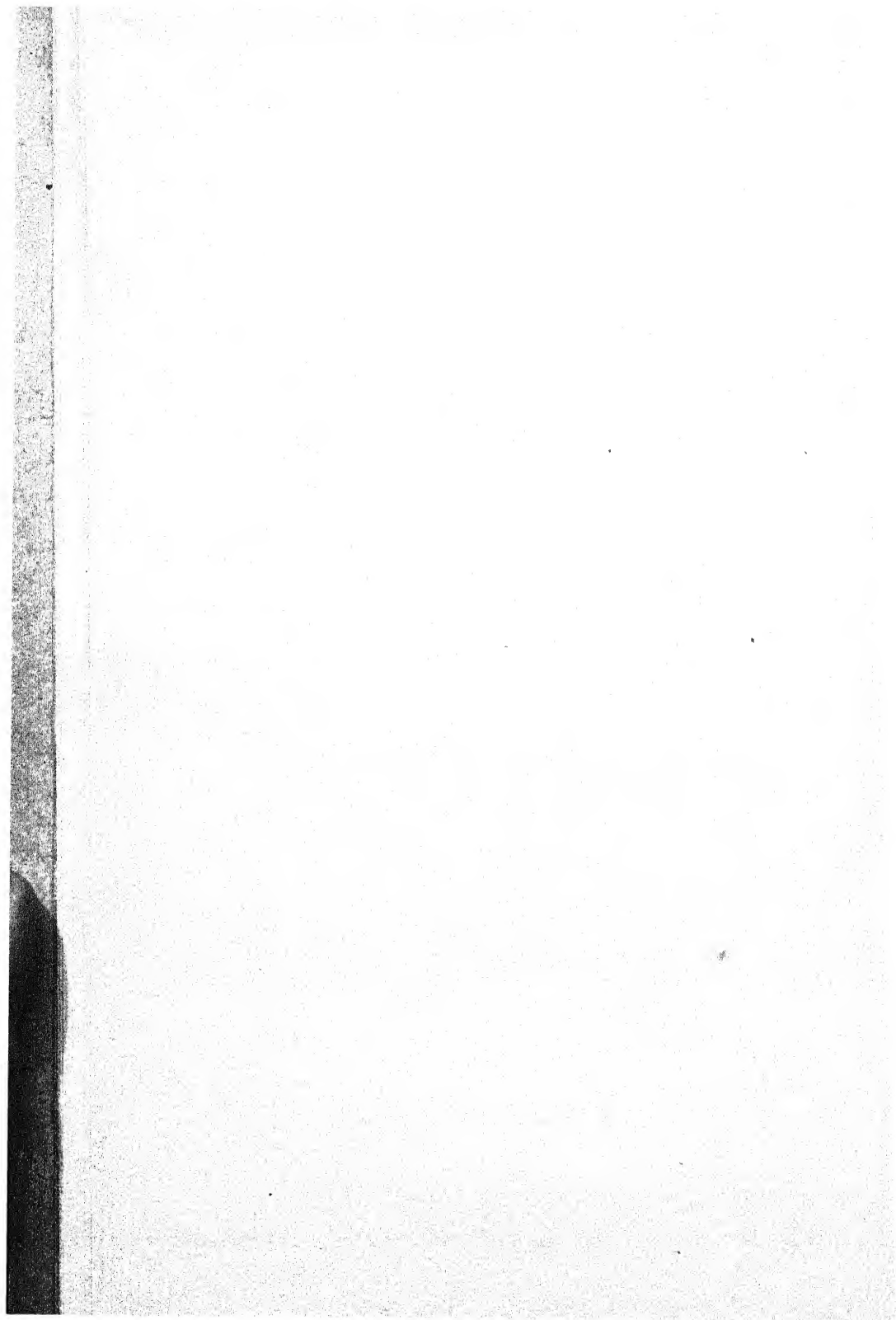
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SOIL PHYSICS

CHAPTER I

INTRODUCTION

The soil is a very complex system. A given volume of soil is made up of solid, liquid and gaseous material. The solid phase may be mineral or organic. The mineral portion consists of particles of varying sizes, shapes and chemical compositions. The organic fraction includes residues in different stages of decomposition as well as live, active organisms. The liquid phase is the soil water which fills part of or all the open spaces between the solid particles and which varies in its chemical composition and the freedom with which it moves. The gaseous or vapor phase occupies that part of the pore space between the soil particles that is not filled with water; its composition may change within rather short intervals of time. The chemical and physical relationships between the solid, liquid and gaseous phases are affected not only by their own respective properties but also by temperature, pressure and light. The mechanical behavior of the soil mass is referred to as the **physical properties** of soils. The study of these properties is known as **soil physics**.

Within the complex soil system grow the plants that are the basis of all agriculture. If the solid phase contains sufficient nutrients that can be released to the plant, the soil is said to be fertile. If the pore spaces between the solid particles are so distributed as to provide ample water storage for plant growth and at the same time permit adequate aeration to plant roots, the soil is considered to have favorable water and air relationships. The ability of the soil to produce crops is dependent not only upon the proper supply of nutrients but also upon such air and water relations as will make possible the most efficient usage of available nutrients by plants. Root development is more extensive and biological and chemical processes take place under optimum conditions. The importance of the physical properties of soils in plant growth has rarely been thoroughly appreciated because of the emphasis that has been placed upon soil chemistry in relation to fertility.

The Work of Schübler. Sir Humphry Davy (1813) was among the first to recognize the significance of the physical properties of soils in agriculture. He was followed by Schübler, who wrote a book, *Grundsätze der Agrikultur-Chemie*, in 1833, in which he reported the first technical investigations in soil physics. In Part II, which is devoted to agronomy, he discussed, along with the chemistry of soils, the "physical properties which influence the productivity of soils." These properties were:

1. The specific gravity and volume weight in the dry and moist states.
2. The water-holding capacity, by weight and by volume.
3. The tenacity and solidity in the dry and wet states.
4. The different capacities to dry out in air.
5. The volume decreases on drying.
6. The absorption of moisture from atmospheric air.
7. The absorption of oxygen from atmospheric air.
8. The heat capacity of soils.
9. The capacity to absorb heat from sunlight.
10. The capacity to develop heat on wetting.
11. The polar-electric behavior and conductivity for electricity.

Schübler studied thirteen various materials, which included different sands, various clays, powdered carbonates and gypsum, humus and productive soils. Several of his results are given in Table 1. It is significant to note that no attempt was made to study the specific effect of particle size. The importance of particle size was recognized, however, since various sand-clay mixtures were compared. Soil structure, moisture, consistency and temperature were given the major emphasis. His work pointed out the value of porosity in the air-water relationships of soils, even though the air capacity was not specifically determined. Expressing the water-holding capacity in terms of percentage by volume and determining volume weight were two important steps in understanding soil porosity. Investigations on water-vapor absorption, heat of wetting, rates of drying and volume changes indicated the need for evaluating the surface properties of soils. These studies were considerably ahead of what was known about clay and absorption. The importance of consistency in the mechanical handling of soils was thoroughly recognized by the work on tenacity, solidity and adhesion. Adhesion measurements were made on wood and iron, since both these materials were then being used for making tillage implements. The significance of color and moisture content on soil-temperature relations was emphasized by the in-

TABLE 1

PHYSICAL PROPERTIES OF VARIOUS SOILS (SCHÜBLER)

Physical property	Type of soil					
	Quartz sand	Light clay: 45 per cent fine sand, 55 per cent clay	Heavy clay: 10 per cent fine sand, 90 per cent clay	Pure clay	Humus	Productive field soil
Specific gravity.....	2.653	2.601	2.560	2.533	1.370	2.401
Volume weight—dry grains per cu. in.....	495	435	357	334	154	376
Water-holding capacity—per cent by weight.....	25	40	61	70	181	52
per cent by volume.....	37.9	51.4	62.9	66.2	69.8	57.3
Tenacity—dry clay = 100.	0	57.3	83.3	100	8.7	33.0
Adhesion—wet on iron, lb. per sq. ft.....	3.8	7.9	17.2	27	8.8	5.8
Capacity to dry out—						
Time to lose 90 per cent H ₂ O at 18.8° C.....	4 hr.	6 hr.	10 hr.	11 hr.	17 hr.	11 hr.
Shrinkage, cc. per 1000....	4 min.	55 min.	19 min.	17 min.	3 min.	15 min.
Water-vapor absorption, gm. per 1000 in 48 hr....	0	60	114	183	200	120
Heat capacity—calcareous sand = 100.....	0	28	40	48	110	23
Temperature of dry surface in the sun—air temp. 25° C.....	95.6	76.9	68.4	66.7	49.0	70.1
	44.8	44.1	44.6	45.0	47.9	44.3

vestigations on heat absorption from direct sunlight. Wet, light-colored soils were always found more difficult to warm up than those which were well drained and dark colored. All these results are outstanding, and the data on specific gravity are cited in many modern textbooks. Schübler's work in soil physics and agricultural chemistry was overshadowed at this time by the work of Liebig and others. Apparently, the enthusiasm for the solution of plant-nutrition and soil-fertility problems through chemistry caused the work in soil physics to be shoved into the background.

The Contributions of Schumacher. The investigations of Schübler were re-emphasized in 1864 by Schumacher in his book *Die Physik*. Most of the soil-physics part of this book was based on Schübler's original data. Schumacher was interested primarily in air and water movement in soils and introduced the concept of capillary and non-capillary porosity as well as the capillary-saturation capacity of soils. The amount of water that a soil could hold, its capillary-saturation

capacity, was considered a function of the size of particle which determined the number and size of the capillary pores. The rapidity with which water moved through the soil was visualized as dependent upon the structure of the soil as it affected the amount of non-capillary pores. Schumacher stressed the importance of the condition of the immediate surface for the entrance of air and water. He also called attention to the effects of the presence of a heavy layer below a permeable surface on the flow of soil water. He discussed water in its relation to agricultural practices and suggested methods of drainage and irrigation that were based upon the physical properties of the soil.

Schumacher attempted to analyze various types of soils from the point of view of their physical behavior and then suggested management methods that would accomplish a physical betterment of the soil. Drainage, deep tillage, liming, addition of sand, use of organic matter and frost action were considered as essential factors for improving clay soils. Irrigation, addition of clay and use of organic matter were recommended for bettering sandy soils. Green manures were regarded as effective means for increasing the capillary capacity of sands. He recognized the importance of the shading effect of plant leaves on the soil and attached much significance to the protection offered by foliage against the beating action of raindrops. The dispersion of soil particles as a result of the impact of the raindrops and the subsequent washing of the particles into the non-capillary pores were visualized as primary causes of soil compaction.

It is difficult to realize how Schumacher's practical analysis of the physical properties of soils in terms of soil behavior and plant relationships could be so completely lost in the enthusiasm of that day for chemical means for solving the farmer's problems. Many of his suggestions, like Schübler's, cannot be improved upon by present-day knowledge.

The Investigations of Wollny. The first extensive investigations on the physical properties of soils were made by Wollny and his associates from 1879 to 1898. During this time Wollny edited a journal, *Forschungen auf dem Gebiete Agrikulturphysik*, which contained original articles and abstracts on the subjects of soil physics, plant physics and agricultural meteorology. Twenty volumes of this journal contain all his investigations as well as those of other workers. His main thesis was built upon the conviction that the growth of crops, as well as fertilizer responses, were primarily determined by the physical properties of the soil. A wide range of related factors was studied. They may be grouped into the following categories:

1. The effect of soil color on temperature, moisture and carbon dioxide relationships.
2. The physical properties of various soil types.
3. The influence of plant cover and shading upon the physical properties of soils.
4. Factors affecting the carbon dioxide content of the soil air.
5. The effect of degree of slope and direction of exposure on soil moisture and temperature.
6. The effect of compaction on the physical properties of soils.
7. Factors affecting the capacity and movement of soil water.
8. Factors affecting soil-temperature relationships.
9. Temperature and moisture relations in different types of forest litter.
10. The effect of rainfall on soils and plants.
11. The effect of soils and plants on the temperature and moisture relationships of the atmosphere.
12. The effect of cultivation on soil productivity.
13. The beneficial effects of earthworms in soils.
14. The importance of the physical properties of soils in the production of crops.

Wollny studied the properties of soils from the point of view of the plant—not only how various soil properties affected plant growth but also how plants influenced the soil. In 1877, he wrote a short monograph, *Untersuchungen über den Einfluss der Pflanzendecke und der Beschattung auf die physikalischen Eigenschaften des Bodens* (Investigations concerning the influence of plant cover and shading upon the physical properties of soils). His investigations on the effect of plant cover on soils, along with the experiments on the effect of rainfall on soils and plants, were pioneering advances in the field of hydrology in its relation to soil physics. As a result of well-planned and clever experimentation, he was able to present a fairly clear picture of the importance of plants and soils in the disposition of precipitated water, or, as expressed in modern terminology, the role of soils and vegetation in the hydrologic cycle. Wollny must have been a keen observer of the various phenomena occurring during a rainstorm to have planned such comprehensive experiments on the disposition of rainfall. His results strongly suggest that he visualized the properties of the soil, in their relation to the associated vegetation, as being the key to effective water conservation.

He measured the interception of rainfall by the foliage of plants and studied the protective effect of the vegetative canopy on soil

porosity. He determined the amount of runoff and erosion that occurred from plots under different degrees and direction of slope, soil and vegetation conditions. Investigations of the influence of soil properties and plant cover on the percolation of water through soils were made by means of lysimeters. These classical experiments produced data that have rarely been cited by most of the present-day soil investigators, even though many of the latest scientific contributions in this field are, in reality, only confirmations of Wollny's original results.

Wollny obtained innumerable data relating to the effect of the physical properties of soils on crop growth and fertilizer responses. Most of these results provided an insight into moisture relationships as they affected soil temperatures, aeration and the availability of water to plants. Much concern was expressed over the poor physical behavior of soils in a compacted condition. Plowing, cultivating and other tillage operations were studied from the standpoint of the soil and the plant. One is led to believe that the major purpose of these studies was to provide information to show that the chemistry of soils was not the answer to the agronomic problems confronting the farmer. This belief is emphasized in one of the latest papers of Wollny in which he attempted to correlate most of his previous work by discussing the importance of the different growth factors on soil productivity.

Owing to the wide scope and fundamental nature of the investigations, the cleverness of the techniques employed and the practical value of the interpretation of his results, Wollny may truly be called the father of soil physics. It is regrettable that his results and practical conclusions have not been more available to soil investigators.

Early Work of American Investigators. During the time of Wollny's studies in Germany, several American investigators were making significant contributions to the knowledge of the physical properties of soils. They were Hilgard of California (4, 5), Johnson of Connecticut (6, 7), King of Wisconsin (8, 9), and Slichter of the United States Geological Survey (11). In 1873, Hilgard developed an elutriator for making silt analyses of soils and subsoils and fractionated soils on the basis of the hydraulic value of the particles. Compactness, which was considered an index of the resistance to tillage, was attributed to those particles having a hydraulic value of 0.25 mm. per second or less (about 0.015 mm. in diameter). Porosity appeared to be correlated with those fractions having a value of 1 mm. per second or more (about 0.03 mm. in diameter).

Johnson, in 1877, discussed the reasons for tillage and suggested that its major objective was to modify water storage in the soil. This was accomplished by giving the crop a better-aerated and healthier

rootbed and by conserving rain against drought. Conservation of water was thought to be achieved through increased absorption of rainfall and decreased vapor losses as a result of less evaporation and weed transpiration. He emphasized the importance of the falling rain-drop in the compaction of the surface soil as a result of the impact of the drop and the clogging of the surface pores by the dispersed silt and clay. He visualized that drying was the real loosening agent in soils. Tillage lifted up the mass of compact soil, turned it over and let it fall so that the close contact between particles disappeared. The individual sand particles and aggregates were thus brought into new positions with regard to each other and placed at greater distances from each other. Hilgard disagreed with Johnson on the mechanical effects of tillage and maintained that these "new positions" and "greater distances" were brought about and maintained by flocculation. It is worth while to note that both men attempted to associate good tilth with granulation, although they pictured the mechanism from different points of view.

In 1878, Johnson performed a series of experiments on the relation of soils to water, primarily from the point of view of capillary transmission and evaporation. He showed that water moved more slowly in dry pores than in moist ones and less rapidly in fine pores than in large ones. In light of recent developments in the movement of soil water, it is important to point out that Johnson characterized the capillary power of the soil as consisting of both distance and speed of movement. The rate of movement was given much greater significance than the distance of travel.

King, who is generally considered the father of soil physics in America, outlined his contemplated investigations in 1888. Most of his work aimed to answer the following questions:

1. What is the capacity of various soils to store water?
2. How much of the water stored in soils can different crops use and to what depth do these crops draw water?
3. What volume of air should soils contain?
4. What is the effect of tillage on soil porosity, temperature and evaporation?

Researches to solve these problems produced the classical data on the water requirements of various crops, the root action of crops in relation to moisture utilization, the conservation of water by means of dust mulches which retard capillary activity and the effect of tillage on soil moisture, aeration and temperature. King apparently visualized the water economy of soils as dependent upon the character-

istics of the plant and the capillary properties of the soil. Conservation of moisture was based upon tillage practices to increase the water capacity of the soil and to prevent the loss of water by evaporation or transpiration through weeds. His theory of dust mulches was based upon evaporation losses from columns of soil in contact with a free-water surface. It served as the fundamental basis of tillage practices for over twenty-five years, but later was found to be not as important as originally suggested. After Slichter (11) had developed a mathematical conception of soil-pore space, as based upon the type of packing of spherical soil particles, King attempted to determine the effective diameter of the pores by means of the rate of flow of air through a given column. From these data it was possible to calculate the effective size of the soil particles according to Slichter's theory.

The results of King's work as soil physicist at the University of Wisconsin were published in 1899 in his book, *Physics of Agriculture*. This book was written from the point of view of the practical value of soil physics and should be inspiring to all present-day workers in this field, particularly his view in relation to soil moisture and tillage problems.

Trends in Soil Physics Since 1900. With the possible exception of investigations in the fields of the soil moisture and mechanical analyses, there was a marked decrease in interest and research on the physical properties of soils after the classical contributions of Wollny and King. In 1900, Warrington of England published *The Physical Properties of Soils*. This book represented the lectures that he gave as professor of rural economy in the University of Oxford in 1896. In America, several workers in the Bureau of Soils of the United States Department of Agriculture made notable contributions to the existing knowledge in soil physics. Briggs (1, 2) was interested in the mechanics of soil moisture and developed the much-used moisture-equivalent technique for characterizing the water relationships of soils. Buckingham (3) initiated the concept of capillary potential and conductivity for understanding the movement of soil moisture. The value of his researches was not fully realized until the recent emphasis on the energy relationships in the movement of soil water. Patten (10) made some rather important findings concerning the transference of heat in soils. His results have served as a basis for most of the textbook discussions in soil temperature. In Europe, some outstanding work was done by Mitscherlich on hygroscopicity and heat of wetting, by Atterberg on mechanical analysis and soil consistency, by Odén on the size distribution of soil particles and by Ehrenberg in the field of soil colloids.

The beginning of the third decade in the present century marked a revival of interest in and thought about the physical properties of soils. Investigators throughout the world began to realize the importance of the mechanical behavior of soils in the study of soil problems. Soil physics began to expand from studies in mechanical analyses to investigations of soil structure, consistency, aeration, water movement, the physical behavior of the soil separates and other fundamental as well as applied aspects of the subject. Numerous contributions to scientific literature have been evidence of the renewed interest in the physical properties of soils. International and national soil-science societies have organized sections of soil physics. Two rather well-known books have been published that discuss the subject, namely, *Physical Properties of Soils* by Keen and *Die physikalische Beschaffenheit des Bodens* (The physical properties of soils), volume 6 of the *Handbuch der Bodenlehre*. There is evidence from the progress made since 1920 that soil physics will soon occupy its rightful place with the other phases of soil science in the solution of soil problems.

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CHAPTER II

THE MECHANICAL COMPOSITION OF SOILS

THE SOIL AS A DISPERSE SYSTEM

The soil is essentially a disperse system, that is, a system in which the particles are in a fine state of subdivision or dispersion. In soils, the dispersed or solid phase predominates, and the dispersion medium, soil water, only helps to fill the pores between the solid particles. Moreover, the solid phase consists of particles of varying degrees of subdivision, which range from the lower limits of the colloidal state to the coarsest fractions of sand and gravel. These different particles, especially those of colloidal dimensions, may be found in a state of almost complete dispersion or in a condition of nearly perfect aggregation, or granulation. In most soils, however, there is only a partial aggregation of the various particles.

The completely dispersed, individual or primary particles are usually referred to as **textural or mechanical separates**. The aggregates or secondary particles, which are formed by a grouping together of mechanical separates, are generally considered the **structural units of the soil**. The properties of the solid phase cannot be evaluated accurately unless both types of particles are considered. This emphasizes the importance of characterizing the type of particles when referring to the physical condition of soils.

The Relation of Surface to Particle Size. The significance of the concept of dispersion becomes more apparent when the relationship of surface to particle size is considered. The large amount of surface per unit mass is a characteristic property of all disperse systems. Colloidal chemistry is primarily surface chemistry or, more correctly, the chemistry of interfaces, since the physical and chemical behavior of colloids depends upon the existence of an interface (two surfaces coming into contact) between two phases of the system. The colloidal state of subdivision of matter is important physically and chemically, owing to the enormous surface per unit mass. The extent of surface of a disperse system is usually expressed in terms of **specific surface**, which signifies the square centimeters of surface per gram or per cubic centimeter of dispersed phase.

It is interesting to follow the increase in surface as a single cube is subdivided into smaller ones. For example, visualize a cube of any material whose sides are 1 cm. long. The volume of such a cube is 1 cc.; the total surface is 6 sq. cm.; and the specific surface is 6 sq. cm. per cc. Let us cut this cube into smaller ones whose sides are $\frac{1}{10}$ as long, namely, $\frac{1}{10}$ cm., or 1 mm. Such particles will be about the size of fine gravel. Each cube will have a volume of $\frac{1}{1000}$ cc. and a surface of $\frac{6}{100}$ sq. cm. But there will be 1000 cubes, which will make the total surface equal to 60 sq. cm. ($1000 \times \frac{6}{100}$). Let us suppose that the original cube is subdivided into cubes whose sides are $\frac{1}{10,000}$ as long as originally, that is, $\frac{1}{10,000}$ cm., or $\frac{1}{1000}$ mm., or 1 μ (micron). These particles will be about the size of fine clay. Such a subdivision will produce 10^{12} cubes with a total surface of 60,000 sq. cm. In order to carry the subdivision of this hypothetical cube into the colloidal state, let us divide it into cubes whose sides are $\frac{1}{100,000}$ as long as they were originally, namely, $\frac{1}{100,000}$ cm., or $\frac{1}{10,000}$ mm., or $\frac{1}{10}$ μ or 100 $m\mu$ (millimicrons). We shall now have 10^{15} cubes with a total surface of 600,000 sq. cm. Thus, it is seen that the specific surface increases from 6 to 600,000 sq. cm. per cc. as the size of the cubes decreases from 1 cm. (gravel) to 100 $m\mu$ (colloidal state) on a side.

Inasmuch as the different textural separates in soils are classified into various groups on the basis of their effective diameters, it is of special interest to know the specific surface of each of these groups. The calculations in Table 2 point out that a very small amount of material (about 0.52 cc.) may exhibit an extraordinarily large surface area if a sufficiently high degree of subdivision is attained. Several interesting comparisons may be made from these results. For example, a given weight, or volume, of 2 μ clay has 50 times more surface than the same quantity of very fine sand; it has 10 times more surface area than the same weight of silt. Colloidal clay (100 $m\mu$) possesses 20 times the surface area of 2 μ clay and 1000 times the surface of very fine sand. These large differences between the amount of surface per unit mass in clay and sandy soils should permit a clearer understanding of the recognized dissimilarities in their physical behavior.

CLAY—THE ACTIVE SOIL FRACTION

Numerous experimental findings have shown that the chemical and physical properties of soil are dependent upon the surface activity of the clay fraction. It has been adequately demonstrated that sand and silt possess little surface activity and, therefore, do not contribute much to many of the physical characteristics of soils. This small sur-

TABLE 2—THE RELATION OF SURFACE TO PARTICLE SIZE

Diameter of sphere	Textural name	Volume per particle ($\frac{1}{6} \pi D^3$)	Number of particles in π cc.	Total surface $\pi D^2 \times$ number of particles
1 cm.....	Gravel	$\frac{1}{6} \pi (1)^3$	1	3.14 sq. cm. = 0.49 sq. in.
0.1 cm.....	Coarse sand	$\frac{1}{6} \pi \left(\frac{1}{10}\right)^3$	1×10^3	31.42 sq. cm. = 4.87 sq. in.
(1 mm.).....				
0.05 cm.....	Medium sand	$\frac{1}{6} \pi \left(\frac{5}{100}\right)^3$	8×10^3	62.83 sq. cm. = 9.74 sq. in.
(0.5 mm. or 500 μ).....				
0.01 cm.....	Very fine sand	$\frac{1}{6} \pi \left(\frac{1}{100}\right)^3$	1×10^6	314.16 sq. cm. = 48.67 sq. in.
(0.1 mm. or 100 μ).....				
0.005 cm.....	Coarse silt	$\frac{1}{6} \pi \left(\frac{5}{1000}\right)^3$	8×10^3	628.32 sq. cm. = 97.34 sq. in.
(0.05 mm. or 50 μ).....				
0.002 cm.....	Silt	$\frac{1}{6} \pi \left(\frac{2}{1000}\right)^3$	125×10^6	1,570.8 sq. cm. = 1.69 sq. ft.
(0.02 mm. or 20 μ).....				
0.0005 cm.....	Fine silt	$\frac{1}{6} \pi \left(\frac{5}{10,000}\right)^3$	8×10^9	6,283.2 sq. cm. = 6.76 sq. ft.
(0.005 mm. or 5 μ).....				
0.0002 cm.....	Clay	$\frac{1}{6} \pi \left(\frac{2}{10,000}\right)^3$	125×10^9	15,708 sq. cm. = 16.9 sq. ft.
(0.002 mm. or 2 μ).....				
0.0001 cm.....	Clay	$\frac{1}{6} \pi \left(\frac{1}{10,000}\right)^3$	1×10^{12}	31,416 sq. cm. = 33.8 sq. ft.
(0.001 mm. or 1 μ).....				
0.00005 cm.....	Clay	$\frac{1}{6} \pi \left(\frac{5}{100,000}\right)^3$	8×10^{12}	62,832 sq. cm. = 67.6 sq. ft.
(0.0005 mm. or 500 m μ).....				
0.00002 cm.....	Colloidal clay	$\frac{1}{6} \pi \left(\frac{2}{100,000}\right)^3$	125×10^{12}	157,080 sq. cm. = 169 sq. ft.
(0.0002 mm. or 200 m μ).....				
0.00001 cm.....	Colloidal clay	$\frac{1}{6} \pi \left(\frac{1}{100,000}\right)^3$	1×10^{15}	314,160 sq. cm. = 338 sq. ft.
(0.0001 mm. or 100 m μ).....				
0.000005 cm.....	Colloidal clay	$\frac{1}{6} \pi \left(\frac{5}{1,000,000}\right)^3$	8×10^{15}	628,320 sq. cm. = 676 sq. ft.
(0.00005 mm. or 50 m μ).....				

face activity is due not only to a relatively low specific surface but also to the chemical and mineralogical nature of these coarser separates; that is, they are primarily quartz or unweathered primary minerals. Owing to the absence of any marked physical or chemical activity, the silt and sand fractions may be considered as the skeleton of the soil. The clay and humus material are the active portions because of their high specific surface and their chemical constitution. Thus, sandy soils are not physically and chemically active because of a large amount of skeleton material. Clays, on the other hand, are known to be plastic and sticky when wet, to shrink on drying and to swell on

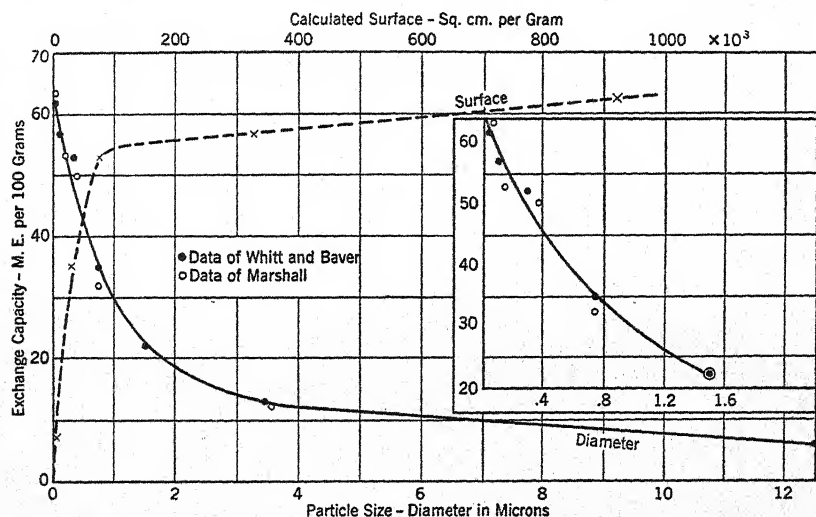


FIG. 1. The relation of particle size and specific surface to the exchange capacity of soils.

wetting; they are poorly aerated and impervious to water. These pronounced physical properties result from the great surface activity of the colloidal particles.)

The important relationship of particle size and specific surface to the properties of clays is illustrated in Figure 1. These data show how the adsorption of exchangeable cations by various soil separates increases as the particle size decreases and the specific surface becomes greater. The experimental findings on the B-horizon of Putnam silt loam from Missouri (35, 54) indicate that coarse silt (20 μ to 5 μ diameter) possesses only about $\frac{1}{10}$ of the surface properties of coarse clay (2 μ to 1 μ diameter) and less than $\frac{1}{20}$ of the adsorptive capacity of colloidal clay (less than 100 $m\mu$ diameter). These results, which

are typical of the data from other related investigations, substantiate the concept that clay is the active soil fraction.

Definition of Clay. The definition of clay has long been debated. The older concepts attempted to characterize clay on the basis of its chemical nature and its effects upon the soil. For example, Osborne (41), who developed the beaker method of mechanical analysis in 1886, defined clay as follows: "By true clay is here meant that material, mostly derived from the decomposition of feldspars and similar silicates, which is capable of uniting with a considerable amount of water, and thus assuming a gelatinous condition in which it exerts a powerful binding action upon the particles of sand in the soil. To some extent, probably, this action is also exerted by iron and alumina hydroxides, as well as by colloid organic bodies."

The "material" to which Osborne referred was first pictured as being kaolinitic in nature; next, it was considered to be similar to zeolites; later, it was defined as an indefinite mixture of the colloidal oxides of silicon, aluminum and iron; finally, as later discussions will point out, it has been shown to consist primarily of definite clay minerals.

The purely chemical definition of clay was replaced by one with a colloidal meaning when Odén (37) and others applied the idea of disperse systems to the study of soils. Odén defined clay as "disperse formations of mineral fragments in which particles of smaller dimensions than $2\ \mu$ predominate." This concept permits clay to consist of primary mineral fragments together with the secondary products of weathering, as long as the individual particle sizes are small enough.

The choice of $2\ \mu$ as the upper limit for clay has an interesting and significant development. Investigators in the Bureau of Soils of the United States Department of Agriculture modified the beaker method of Osborne in 1896. They extended the separation of the smallest particles from 0.1 to 0.005 mm. and gave the latter limit the designation of clay. The choices of the different limits were arbitrarily made apparently on the basis of convenience of calibration with the particular eyepiece micrometer that was used. This is illustrated in the following statements: "With the microscope used in this Division the 1 inch eyepiece and $\frac{3}{4}$ inch objective, three of the 0.1 mm. spaces of the eyepiece micrometer measure 0.05 mm. on the stage. With the same eyepiece and the $\frac{1}{2}$ inch objective, two spaces of the micrometer are equal to 0.01 mm., and one space to 0.005 mm. These three values are sufficient for the beaker separations (53)." The resulting classification of soil separates, which is given in Figure 2, was used in the

United States until 1937. In 1912, Atterberg (2) suggested $2\ \mu$ as the upper limit of clay. His classification of soil particles (Figure 2) according to size was accepted by the International Society of Soil Science in 1913 and was adopted in the Bureau of Chemistry and Soils in 1937 to replace eventually the modified Osborne groupings.

Atterberg attempted to give some scientific basis for setting up the various size limits and for characterizing clay as those particles smaller than $2\ \mu$ (0.002 mm.). The first group, 20 to 2 mm., was placed between the limits where no water is held in the pore spaces between the particles and where water is weakly held in the pores. The lower limit of the second group, 2 to 0.2 mm., was set at the point where water is held in the pores by the forces of capillary attraction. The lower limit of the third group, 0.2 to 0.02 mm., was given the theoretical

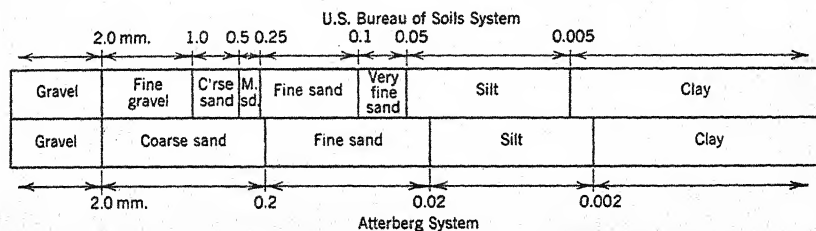


FIG. 2. A Comparison of the Atterberg and old Bureau of Soils systems of particle-size separations (Truog, *et al.*, 51).

significance that smaller particles cannot be seen by the naked eye, do not have the usual properties of sand and can be coagulated to form the "crumbs" that are so significant in the mechanical handling of soils. In other words, these are the limits between dry sand, which gives poor soils, and adequately moist sand, which forms productive sandy soils. The lower limit of the fourth group, 0.02 to 0.002 mm., was established on the basis that particles smaller than $2\ \mu$ (clay) exhibited Brownian movement in aqueous suspension. Capillary movement of water is very slow for particles of this size, and the properties of stiff clays are strongly manifested. Thus, silt was visualized as a range of particle sizes from the point where sand begins to assume some clay-like properties to the upper limit of clay.

Atterberg's definition of clay has been given further scientific justification as a result of recent mineralogical studies of soils. Different investigators have noticed changes in the composition and properties of soil particles in the region of $2\ \mu$. Robinson (46) has considered that the material below 0.002 mm. is primarily the colloidal products of weathering and is the truly active portion of the

soil from the chemical point of view. Recent investigations of Marshall (34), as well as of Truog and his associates (51), have indicated that very few unweathered primary minerals exist in the fractions below $2\ \mu$ in diameter. This is particularly true of the feldspars, as shown in Table 3. Therefore, in light of these studies, Odén's

TABLE 3
MINERALOGICAL COMPOSITION OF VARIOUS SOIL FRACTIONS

Soil	Minerals in separates indicated (35)			
	$20\ \mu - 5\ \mu$	$5\ \mu - 2\ \mu$	$2\ \mu - 1\ \mu$	$1\ \mu - 500\ m\mu$
Putnam clay	feldspars muscovite quartz	feldspars muscovite	quartz intermediate material between feldspars and clay minerals	clay minerals primarily with some intermediate material
Rothamsted clay	feldspars muscovite	quartz muscovites chlorites (?)	quartz clay minerals and intermediate material	clay minerals primarily
Coal measures clay	quartz feldspars glauconite halloysite	quartz feldspars clay minerals chlorites (?)	clay minerals primarily	clay minerals primarily
	Abundance of feldspars in separates indicated (51)			
	$10 - 5\ \mu$	$5 - 4\ \mu$	$4 - 3\ \mu$	$3 - 2\ \mu$
Miami silt loam	abundant	moderately abundant	moderately abundant	trace
Carrington silt loam	abundant	moderately abundant	moderately abundant	trace
Colby silt loam	abundant	moderately abundant	present but not abundant	trace
Hagerstown silt loam	abundant	moderately abundant	moderately abundant	trace
Cecil clay	present but not abundant	present but not abundant	not detected	not detected

definition of clay should be modified to read [clays are disperse systems of the colloidal products of weathering in which secondary mineral particles of smaller dimensions than $2\ \mu$ predominate.]

Some attempt has been made to distinguish between clay and colloidal clay. The Bureau of Chemistry and Soils has attached considerable significance to the particles smaller than $1\ \mu$ (0.001 mm.) which it has considered to be colloidal in nature (10). Bray (9) has also considered $1\ \mu$ as the upper limit of colloidal clay. Recent studies on the properties of various fractions from Putnam silt loam (54) have suggested that $1\ \mu$ is not far from the upper limit of colloidal size in this soil if this boundary is taken, first, as that point below which physical and chemical properties are nearly constant and, second, where a distinct break occurs in the curve showing the different properties of the clay as a function of particle size. This limit is considerably larger than that suggested by colloid chemists for the maximum size of colloidal particles. Although Freundlich (15) has set $0.5\ \mu$ ($500\ m\mu$) as the upper limit of the colloidal range, most colloid chemists have considered $200\ m\mu$ as the maximum size. It may be possible to reconcile these differences on the basis of the large intermicellar surface in the clay minerals which would make the large clay particles exhibit colloidal properties.

On the other hand, the investigations of Joseph (23) and DeYoung (13) have indicated that all the particles smaller than $2\ \mu$ may be reduced to the colloidal state if the sample is completely dispersed; in other words, the clay and colloid contents of a soil are identical. This concept has resulted from the experimental experience that all a dilute clay suspension apparently could be made to pass through a supercentrifuge if the process of redispersion and recentrifuging were continued long enough. There is some doubt as to the validity of this concept in light of more recent studies in which it was found that the various separates below $2\ \mu$ in size were stable under rather drastic dispersion treatments (54).

On the basis of existing information it seems satisfactory to use $2\ \mu$ as the upper limit of clay. However, if investigations on the colloidal fraction are to be made, it is highly probable that the most reliable results will be obtained if $100\ m\mu$ to $200\ m\mu$ are used as the upper limit of the colloidal range. Such colloidal material not only will conform more nearly to the accepted standards in colloidal chemistry but also will possess a much greater chemical and physical activity per unit weight than the coarser fractions.

The Chemical and Mineralogical Nature of Clay. Although it is not the purpose of this book to discuss thoroughly the chemical and mineralogical nature of clay, it is necessary to have a general idea of the constitution of clay in order to understand its physical behavior. A chemical analysis of the clay fraction of soils usually shows that it is

composed primarily of SiO_2 , Al_2O_3 , Fe_2O_3 and H_2O along with varying amounts of MgO , CaO , K_2O , Na_2O and P_2O_5 . The question naturally arises as to the types of chemical compounds or substances in the soil that may contain these various elements and exhibit the colloidal properties of clays. Is clay a mixture of the hydrated oxides of silicon, aluminum and iron, or is it a distinct substance in which these elements are chemically united? Prior to the work of Bradfield (8), there was considerable argument in favor of the theory of mixed hydrated oxides of silicon, aluminum and iron, with the bases present in the adsorbed state. Bradfield has shown, however, that clay does not have the properties of colloidal SiO_2 , $\text{Al}(\text{OH})_3$ and $\text{Fe}(\text{OH})_3$ when mixed in the proportions in which a chemical analysis of clay indicates they might exist. He has presented sufficient evidence to prove that clay has the properties of a weak acid when saturated with the hydrogen ion, and has suggested that colloidal clay is constituted of complex aluminosilicic acids.)

The application of X-ray and petrographic techniques to mineralogical studies of clays by Ross and Shannon (47), Hendricks and Fry (19), Pauling (42), Kelley, Dore and Brown (27), Hofmann, Endell and Wilm (20), Marshall (34) and others has led to a clearer concept of the nature of clay. These investigators have proved that clays are constituted of distinctly crystalline minerals. Pauling, for example, has shown that the micas and other clay minerals are built up of units of alumina and silica. The alumina unit consists of two sheets of closely packed oxygen atoms or hydroxyl groups which are held together by aluminum atoms in such a way that one aluminum is surrounded by six oxygen atoms, or hydroxyl groups, three from each sheet (Figure 3a). The silica unit consists of a sheet of oxygen atoms that are held together by silicon atoms. The silicon atoms are surrounded by four oxygen atoms, three in the sheet and one above (Figure 3a). The oxygen atom above the sheet provides the means for linking a silica unit to an alumina, since it can serve as one of the six oxygen atoms which surround the aluminum in the alumina sheet.

Two groups of clay minerals have been recognized. These are the kaolin group, or 1:1 type of crystal lattice, and the montmorillonite group, or 2:1 type of lattice. The kaolin minerals include halloysite, metahalloysite, kaolinite, nacrite, dickite and others, and are built up of one silica sheet to one alumina (Figure 3b). The montmorillonite group includes pyrophyllite, montmorillonite, beidellite, nontronite, hydrous mica and others in which the crystal lattice is composed of units of two silica sheets to one alumina (Figure 3b).

Hofmann, Endell and Wilm have shown that the crystal lattice of montmorillonite and beidellite expands and contracts with the amount of water that is present. These results have pointed to a large internal or intramolecular surface. Other investigators have confirmed these findings. Gruner (18) has found that nontronite also has an expanding lattice.

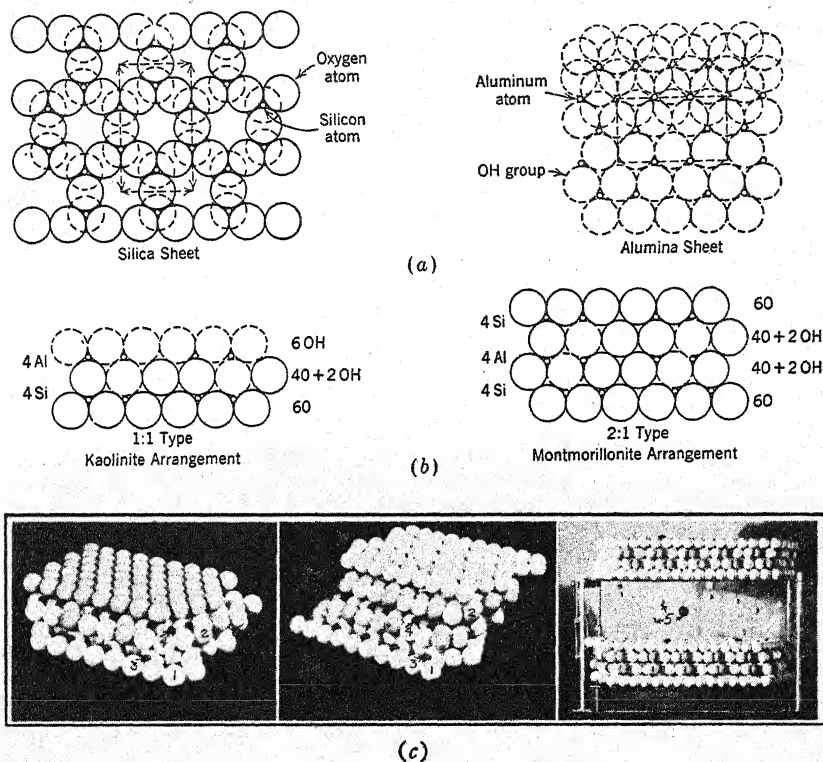


FIG. 3. Crystal-lattice makeup of colloidal clays. (Courtesy of C. E. Marshall.)

The physical and chemical properties of these two groups of clay minerals are rather dissimilar as a result of the differences in their crystal-lattice makeup. The kaolin minerals, with fixed crystal lattices, exhibit only slight hydration and adsorptive properties. Base exchange, or cation adsorption, is very low. On the other hand, the expanding lattices of montmorillonite and beidellite are responsible for high hydration and cation adsorption. Cations and water are adsorbed not only on the outer surfaces but also on the internal surfaces within the crystal. The latter type of surface is undoubtedly

of greater magnitude than the former ; most of the exchangeable cations are found between the sheets as suggested in Figure 3c.

Although clay is composed primarily of definite secondary minerals, varying amounts of the free oxides of iron, alumina and silica may also be present. Usually, they represent but a small portion of the total clay content. Their presence, however, may affect the results of detailed mineralogical analyses. Truog and associates (51) have developed a special technique for their removal prior to making mineralogical studies.

DETERMINATION OF THE MECHANICAL COMPOSITION OF SOILS

The purpose of determining the mechanical composition of soils (mechanical analysis) is to obtain the percentage distribution of the various sizes of individual particles within the soil. The success of such an analysis depends upon (1) the preparation of the sample to insure the complete dispersion of all aggregates into their individual primary particles without breaking up the particles themselves and (2) the accurate fractionation of the sample into the various separates.

PREPARATION OF THE SAMPLE

The main objectives in the preliminary treatment of the soil sample preparatory to a mechanical analysis are to obtain maximum dispersion and to maintain this dispersion during the process of making the analysis. Therefore, it is essential to know the different factors that determine the effectiveness of dispersion and to employ methods that will achieve the maximum results.

Factors Affecting Dispersion. **CEMENTING AGENTS.** Secondary soil particles, or aggregates, are formed by the union of primary particles into units of varying stability. These aggregates may be held together rather loosely or they may be resistant to dispersion because of the presence of so-called cementing agents. The most important of these cementing materials are organic matter, colloidal clay and the dehydrated colloidal oxides of iron and alumina. The significance of these agents in the genesis of soil aggregation will be discussed in detail in Chapter V. It is sufficient to keep in mind at this time that a high degree of correlation exists between the aggregation of clay and silt particles and the amounts of clay and organic matter present, with the exception of lateritic soils. The dehydrated oxides of iron and alumina are primarily responsible for the stable aggregation of lateritic soils.

Robinson (44) was among the first to show that soils containing appreciable quantities of organic matter cannot be adequately dispersed unless the organic material is removed by oxidation with hydrogen peroxide. Fourfold increases in the percentage of clay were obtained by treating soils with hydrogen peroxide. Other investigators also have found that oxidation of organic matter with hydrogen peroxide is essential for the complete dispersion of soil particles. Boiling soils with 6 per cent hydrogen peroxide is a generally accepted procedure in most mechanical-analysis determinations and has been officially adopted by the International Society of Soil Science. This treatment is highly effective and produces only carbon dioxide and water as the by-products of oxidation.

The removal of the cementing influences of the oxides of iron and alumina is a more difficult problem. Although there is little doubt concerning the desirability of removing their binding effects on soil aggregates through adequate dispersion techniques, there is considerable questioning of the validity of removing them from the soil suspension through solution processes. They are a part of the soil colloidal fraction even though they are present presumably as the free oxides. They constitute a significant portion of the inorganic colloidal material in lateritic soils. Truog and associates (51) have suggested the complete removal of these oxides for certain types of mechanical analyses and mineralogical studies, by the use of an oxalic acid-sodium sulfide treatment. Such a treatment reduces the iron to the ferrous state and accomplishes complete solution of the iron and aluminum oxides. This technique makes possible the cleaning up of the various separates so that discrete mineral particles are obtained. It undoubtedly has many possibilities for special studies but should be used with discretion in ordinary mechanical analyses.

REHYDRATION OF CLAY PARTICLES. When soils are fairly moist, the clay particles are surrounded by a water film. As dehydration takes place these films become thinner and thinner until adjacent particles are held together by strong cohesive forces. Dehydrated clay colloids, therefore, can exert a vigorous cementation action on the soil aggregates. The reversibility of these dehydrated secondary particles may be very slow, especially if the colloid content is large. The dispersion of such aggregates requires a rehydration of the clay so that every particle is again surrounded by a water film. This can be accomplished by mechanical agitation in water, such as boiling, shaking, trituration or stirring.

Puri and Keen (43) have shown very convincingly how the original moisture content of the soil and the time of shaking affect the

degree of dispersion. Their results are illustrated in Figure 4. There are several important relationships exhibited by these curves. First, the degree of dispersion varies with the moisture content. At the end of 24 hours of shaking the dispersion factor of the soil with 22.6 per cent of water is more than twice as great as that of the air-dried soil with 3.62 per cent.

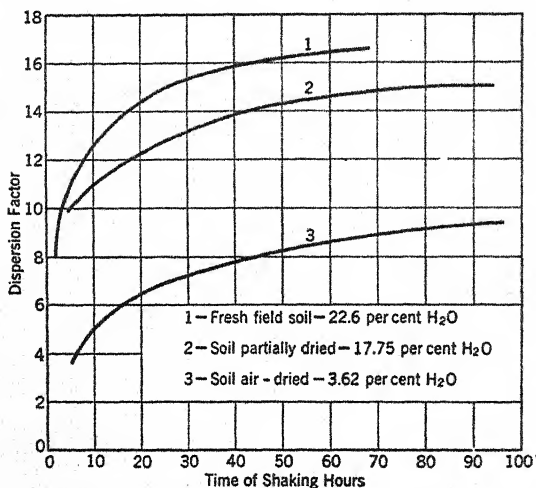


FIG. 4. The effect of soil moisture and time of shaking on dispersion (Puri and Keen, 43).

since there is less cementation. Third, after 20 to 24 hours, the increase in dispersion is very slow and the curves are almost parallel.

Puri and Keen have found that these curves can be evaluated satisfactorily by the formula:

$$d = a + K \log t \quad (1)$$

where d is the dispersion factor, t is the time of shaking and a and K are constants. The value of a varies with the amount of readily dispersible material. The value of K is nearly the same for each of these curves. These results, along with the findings of other investigators, emphasize the importance of the hydration factor in the dispersion of clays.

THE EFFECT OF CHARGE. Removal of the cementation effects of organic matter and other colloidal materials and the formation of a water film around each clay particle are not sufficient to insure maximum dispersion. Particles may be separated physically in a suspension but they may coalesce again into floccules or aggregates before

an accurate mechanical analysis can be made. The magnitude of the electric charge (electric potential) is the determining factor in the stability of a clay suspension. Most clay particles carry a negative charge. If placed between two electrodes of unlike sign they migrate to the positive pole. The greater the speed of migration (migration velocity), the higher is the negative potential. If this potential is sufficiently great, the clay particles will repel each other when they collide during their constant haphazard movement (Brownian movement) in the suspension. If the electric potential is small, there will be little tendency for repulsion and the particles will coalesce as a result of a collision and settle out as a floccule. The effect of charge may be compared with the observations in the classical pith-ball experiment in elementary physics which illustrates that like charges repel and unlike charges attract.

The question arises as to the origin of the charge on clay particles. This should not be difficult to understand in light of the previous evidence which has been pre-

sented to show that clays are primarily aluminosilicate minerals. Adsorbed cations which are capable of replacement are found on the surface of the clay minerals. Therefore, clay particles are essentially colloidal anions with the corresponding cations adsorbed on the surface. Even as sodium acetate may be visualized as CH_3COONa , so may a sodium-saturated clay be pictured as clay-Na, or Na-clay. One colloidal-clay particle, however, will adsorb many cations. Consequently, it may be conceived as a negatively charged nucleus surrounded by positively charged ions, as shown schematically in Figure 5a. It is readily seen that there is an electrical double layer at the surface of the colloidal particle. The colloidal chemist knows this electrical arrangement as the Helmholtz double layer. The inner layer is part of the wall of the particle. It migrates with the particle in an electric field and determines the sign of the charge on the particle; in the illustration given, the charge is negative. The outer layer is of opposite sign and is at a distance of molecular dimensions from the inner. The ions in the outer layer are easily replaced by other ions in the process of ionic exchange.

The magnitude of the electric potential of clay particles will de-

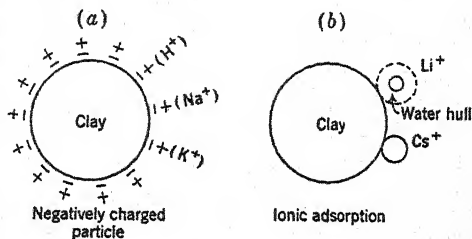


FIG. 5. Schematic representation of colloidal-clay particles in relation to charge (a) and ionic adsorption (b).

pend upon the number of charges per unit surface and the degree of activity of these charges. In other words, the stability of various clay suspensions increases directly with the cation exchange capacity of the clay and inversely with the energy with which the adsorbed cations are held against the inner layer. For a given type of clay particle, the stability will vary with the nature of the adsorbed ion. Wiegner (56) was one of the first to point out that the degree of dispersion, or stability, of clay is determined by the adsorbed cations. He showed that the stability of clay particles, the surfaces of which are saturated with different cations, follows the Hofmeister ion series: $\text{Li} > \text{Na} > \text{K} > \text{Rb} > \text{Cs}$. Clays are more strongly hydrated and more stable, the higher the hydration of the adsorbed cation. Clays saturated with divalent cations are less stable than those containing monovalent ions in the outer layer.

Marshall (33) has carried out mechanical analyses of bentonite and clays and has observed that the nature of the clay mineral as well as the exchangeable cations affect the dispersion. The dispersion of bentonite, which is primarily composed of montmorillonite, is about the same for Mg, Ca, Ba, H, Li, Na, K and NH_4 ions. This similarity in the dispersion is characteristic of montmorillonitic clays and is probably associated with the high hydration that exists in the presence of all cations. On the other hand, Putnam and Rothamsted clays, which contain beidellite, show distinct differences in the degree of dispersion in relation to the type of adsorbed ion. Mg-, Ca-, Ba-, and H-clays are not as highly dispersed as those saturated with Li, Na, K and NH_4 ions.

Jenny (22), in his investigations on the mechanism of ionic exchange in colloidal clays, has established a definite relationship between ionic exchange and the stability of clay suspensions. He has shown that the energy with which various cations are held on the surface of the clay is determined by the effective size of the ion. The energy of adsorption determines the potential of the particle and the stability of the suspension.

According to Coulomb's law, the force of attraction between two ions in a given dielectric varies inversely as the square of the distance between them:

$$F = k \frac{e_a \cdot e_c}{d^2} \quad (2)$$

where e_a and e_c are the charges on the anion and cation, respectively, and d is the distance between them. On a colloidal-clay particle, the

force of attraction between an adsorbed cation and the anion in the negatively charged inner layer will be:

$$F = k \frac{e_a \cdot e_c}{(r_a + r_c)^2} \quad (3)$$

where r_a and r_c are the radii of the anion and cation, respectively. This equation shows that the force of attraction will increase as the radius of the adsorbed cation decreases.

Experimental results have indicated, however, that ionic exchange apparently acts opposite to the requirements of Coulomb's law. For example, the small Li ion is held with much less energy than the large Cs ion. This apparent reversal of Coulomb's law is explained on the basis of the hydration of ions. Ions in solution have a larger effective radius than in a crystal owing to the fact that a water hull is present around the ion in solution. The attraction of water by ions is essentially electrostatic in nature. Water molecules are dipoles and are attracted by the ion, causing an orientation of the molecule with its negative end towards the cation. Since the electric field intensity of an ion increases directly with its charge and inversely with its radius, small ions attract the largest number of water molecules and have the thickest water hull. The crystal lattice and hydrodynamic radii of several ions are presented in Table 4. A schematic picture of ionic hydration in relation to adsorption is shown in Figure 5b.

According to this explanation, the highly hydrated Li ion is larger than the weakly hydrated Cs ion and, therefore, cannot approach the negative inner layer of the clay as closely as the latter. The force of attraction is weak and the ion is easily displaced by others that have greater energies of adsorption. The Helmholtz concept of the electric double layer around a colloidal particle states that the potential is higher the greater the number of electric charges on the surface and the larger the average distance of the outer layer from the inner. It is obvious that hydrated cations in the outer layer will increase the distance of the double layer and cause rather high potentials.

The data of Jenny and Reitemeier (22) on ionic hydration and the potential (migration velocity) of clay particles are given in Table 4 and Figure 6. It should be noted that divalent cations of the same size as a corresponding monovalent ion decrease the potential and, therefore, lower the stability of the suspension. H-clay has a lower migration velocity than any clay system saturated with divalent ions. These data illustrate the extreme importance of charge in the dispersion of soils and call attention to the necessity of substituting a highly hydrated

TABLE 4
DATA ON IONIC SIZE, ADSORPTION, CHARGE AND FLOCCULATION
(JENNY AND REITEMEIER, 22)

Ion saturating the colloidal clay	Size of ion—radii in Angstroms *		Release of adsorbed ions (symmetry values †)	Migration velocity (all negative)	Flocculation value with KCl
	Dehydrated	Hydrated			
	Å	Å	Per cent	μ/sec./volt/cm.	‡
Li ⁺	0.78	10.03	68.0	3.45	21.6
Na ⁺	0.98	7.90	66.5	3.31	11.2
K ⁺	1.33	5.32	48.7	7.8
NH ₄ ⁺	1.43	5.37	50.0	3.48	5.4
Rb ⁺	1.49	5.09	37.4	3.25
Cs ⁺	1.65	5.05	31.2	3.02	5.6
H ⁺	?	?	14.5	2.84	1.5
Mg ⁺⁺	0.78	31.32	3.18	2.9
Ca ⁺⁺	1.06	28.80	3.27	3.0
Sr ⁺⁺	1.27	25.76	3.06	2.6
Ba ⁺⁺	1.43	26.75	3.01	2.3
La ⁺⁺⁺	1.22	13.96	2.74	0.86
Th ⁺⁺⁺⁺	1.85	3.11	0.60

* Angstrom unit, Å = 10⁻⁸ cm.

† Symmetry value = per cent release of an adsorbed ion when replacing ion is added in equivalent amounts to the ion adsorbed. NH₄Cl used with monovalent ions; KCl used for polyvalent ions and NH₄⁺.

‡ Values refer to the number of times symmetry concentration.

monovalent cation for H, Ca, Mg and other strongly adsorbed ions on the surface of clay particles before maximum dispersion can be obtained. Flocculation will be discussed in more detail in Chapter III.

Methods for Obtaining Dispersion. PHYSICAL PROCEDURE. Various techniques have been used to effect the mechanical disruption of aggregates into completely dispersed particles. Among the more important are shaking, stirring, boiling, rubbing or trituration and washing.

Shaking in water or in a peptizing solution is one of the most widely used methods for obtaining adequate dispersion. Reciprocating, end-over-end or rotary types of shakers are most commonly employed. The time of shaking varies with different investigators, ranging from 1 to 24 hours. The experiments of Puri and Keep (43) have shown that dispersion increases with the time of shaking (rotary type) but that the increase in dispersion is very low after 20 to 24 hours. From

the standpoint of routine analysis, however, the convenience of allowing the shaking apparatus to run overnight suggests that 12 to 16 hours will be the maximum time of shaking under most conditions. Shorter times have been used with considerable success, especially if the reciprocating type of shaker is employed. Joseph and Snow (24), for

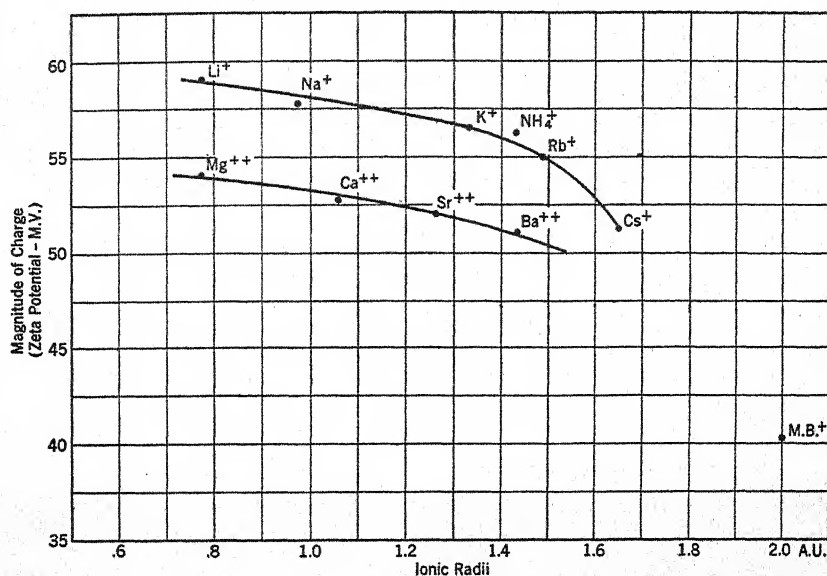


Fig. 6. The relation of the magnitude of the charge on clay particles to the radius of the adsorbed ion (Jenny and Reitemeier, 22).

example, have found that 2 hours are sufficient to disperse heavy alkaline soils.

Stirring with high-speed mechanical stirrers is a rapid method for achieving thorough dispersion. The development of the possibilities of the electric drink-mixing machine by Bouyoucos (4) has made possible rather complete dispersion in a few minutes. He has found that dispersing a soil 10 minutes with the mechanical stirrer is more effective than 16 hours of shaking in a reciprocating shaker. Rapidity is the chief asset of this method. This is especially important when a mechanical analysis is desired on a large number of samples.

Boiling has been employed as a simple means of accomplishing dispersion. Wiegner (57) found that boiling for 1 hour is as effective as shaking for 6 hours. The effectiveness of boiling depends upon the electrolyte content of the suspension. If the soil suspension is previously washed to remove electrolytes, boiling causes a high degree

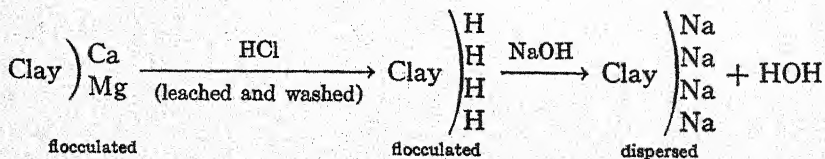
of dispersion. If salts are present, however, boiling increases the coagulating effects of the electrolyte and results in decreased dispersion. These facts should be kept in mind if boiling is chosen as the mechanical means for dispersing soils. It is worth while to note in this connection that the heating of soils during the oxidation of organic matter with hydrogen peroxide materially aids dispersion during any subsequent mechanical treatments.

Trituration is accomplished by gently rubbing the sample in water in a mortar with a rubber pestle. The dispersed material is poured into a beaker and the process repeated until all the smaller particles have been dispersed and removed. It is a time-consuming operation and finds its greatest merit with small samples that require special techniques in their handling.

Wiegner (57) showed that washing electrolytes from soils is essential to obtain a high degree of dispersion by any dispersion method. This is particularly true of soils from regions of limited rainfall where water-soluble salts tend to accumulate. In humid regions, however, the amount of soluble electrolytes in the soil is usually not large enough to cause flocculation if a good dispersion agent is added.

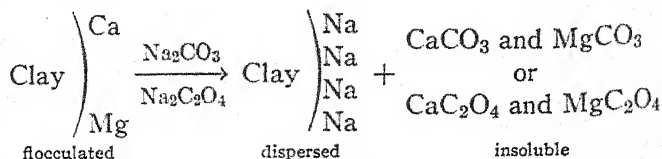
CHEMICAL PROCEDURES. Chemical methods as aids in soil dispersion are based on three distinct principles, namely: oxidation of organic matter, removal of flocculating ions and the peptization of the colloidal particles through the introduction of ions that increase the negative potential. Oxidation of organic matter is accomplished with the use of hydrogen peroxide, as previously discussed.

Removal of flocculating ions, particularly strongly adsorbed cations, is effected in several ways. If divalent basic ions such as calcium or magnesium are present, they may be removed by leaching with dilute acids, by electrodialysis or by precipitation in an insoluble form. Leaching with acids or electrodialyzing the soil sample substitutes the strongly held hydrogen ion for the basic cations on the surface of the particles. After the sample is washed free of electrolytes, hydrogen must be replaced by a peptizing cation, such as Li, Na or NH_4 . This process may be represented as follows:



The success of this technique depends upon the complete removal of chlorides after leaching with acid.

Calcium and magnesium ions may be rendered ineffective also by precipitation as insoluble compounds, such as carbonates or oxalates. This is accomplished by adding sodium carbonate or sodium oxalate to the soil sample prior to shaking or stirring. This treatment causes a replacement of the adsorbed divalent cations by the highly dispersive sodium ion and the precipitation of the calcium or magnesium as carbonates or oxalates:



If NaOH is used with soils containing many divalent cations, the $\text{Ca}(\text{OH})_2$ that is formed as a result of base exchange will exert a strong flocculating action and reduce the degree of dispersion.

Olmstead, Alexander and Middleton (40) have observed that sodium oxalate is more effective than the carbonate in causing maximum dispersion; sodium carbonate causes greater dispersion than the hydroxide. (See Table 5.) Alexander (1) has shown that sodium oxalate may not give as complete dispersion as sodium hydroxide with soils that contain considerable sesquioxides (Fe_2O_3 and Al_2O_3). Green (17) has pointed out that high carbonate parent materials are difficult to disperse with ordinary deflocculating agents. Complete removal of carbonate particles by prolonged acid treatment is not desirable with these soils. He has found that aspirating the sample with CO_2 to convert a part of the relatively insoluble CaCO_3 to $\text{Ca}(\text{HCO}_3)_2$ and subsequent removing of the soluble salts by washing result in satisfactory dispersion. Good results are also obtained with a dilute HCl treatment to remove the more soluble carbonates.

The fundamental basis for the efficiency of various peptizing agents lies in their ability to increase the negative potential of the clay particles. Since this effect increases with the hydration of the ion, it should be expected that the dispersive power of the common monovalent cations would vary according to series: $\text{Li} > \text{Na} > \text{K}$ and NH_4 . Ammonium hydroxide was used for a long time as the standard peptizing agent. It has the advantage that it can be removed from the sample on heating. Later experiments, however, have proved that it is not as effective as sodium hydroxide. Lithium carbonate has been shown to be more effective than sodium carbonate, but its cost makes its use not very practical. In light of this fact, sodium compounds are used most extensively as peptizing agents. Sodium hydroxide is

TABLE 5
THE DISPERSION OF SOILS BY VARIOUS METHODS

Soil	Method of treatment	Per cent clay	Investigator
Stockton clay adobe.....	Shaken 7 hr. with 100 cc. H_2O + 1 cc. conc. NH_4OH .	38.3	Davis and Middleton (12)
	Rubbing and decantation.....	49.8	" "
	Shaken 2 hr. with 100 cc. 0.2 per cent Na_2CO_3	54.0	" "
	Boiled 1 hr. in 0.1 N HCl , washed, shaken 7 hr. in 0.05 N $NaOH$	57.5	" "
5.4 per cent organic matter.	Original.....	28.7	Robinson, G. W. (44)
	Treated with H_2O_2	32.0	" "
11.0 per cent organic matter	Original.....	3.3	" "
	Treated with H_2O_2	13.0	" "
Wabash silt loam less than 2 per cent $CaCO_3$	Shaken 16 hr. with NH_4OH	30.7	Olmstead, Alexander, Middleton (40)
	" " " $NaOH$	33.2	" "
	" " " Na_2CO_3	33.1	" "
	" " " $Na_2C_2O_4$	34.5	" "
Houston black clay 13.9 per cent $CaCO_3$	Shaken 16 hr. with NH_4OH	38.4	" "
	" " " $NaOH$	44.5	" "
	" " " Na_2CO_3	61.3	" "
	" " " $Na_2C_2O_4$	63.8	" "
Cecil sandy loam.....	Shaken 16 hr. with $NaOH$	58.6	Alexander (1)
	" " " $Na_2C_2O_4$	48.9	" "
Davidson clay loam.....	Shaken 16 hr. with $NaOH$	54.0	" "
	" " " $Na_2C_2O_4$	45.0	" "

used with H-saturated systems; sodium carbonate, sodium oxalate and sodium silicate are employed with soils containing exchangeable or carbonate calcium.

Summary on Dispersion. The data in Table 5 point out that a combination of dispersion techniques is necessary to obtain the maximum amount of clay in suspension. It is obvious that the removal of cementing and flocculating agents, vigorous mechanical stirring or shaking and the addition of an effective peptizing agent are all essential. In recognition of this fact, the International Society of Soil Science has adopted the following method of dispersion: Boil the sample with 6 per cent hydrogen peroxide to oxidize the organic matter, treat with 0.2 *N* HCl, wash to remove dissolved carbonates, as well as replaced divalent cations, and shake in 0.008 *N* NaOH to peptize the clay.

Modifications of this procedure are used in nearly all mechanical analyses at present. If small amounts of organic matter are present, the peroxide treatment is often eliminated. Most of the organic colloids will rehydrate and go into solution in the presence of sodium ions. If small quantities of divalent cations are present, the treatment with HCl is not carried out, but these ions are precipitated by adding sodium oxalate to the sample. In the majority of cases the type of dispersion procedure to use will depend upon the fundamental objective of the analysis.

FRACTIONATION OF THE SAMPLE

Inasmuch as soil consists of particles of various sizes and since the fundamental objective of a mechanical analysis is to determine the percentage distribution of these groups of various sizes in the soil mass, the question naturally arises as to the means of accomplishing such an analysis. If a nest of graded sieves were available, separation into different fractions might be accomplished. It is immediately obvious, however, that only the coarser fractions could be separated in this manner. A different principle must be used with the smaller particles. It has been shown that the rate of fall of particles in a viscous medium depends upon the size, density and shape of the particle. In a given medium, such as water, larger particles fall more rapidly than smaller ones with the same density and, consequently, settle out of suspension more quickly. This principle serves as the basis of practically all mechanical analyses.

Fall of Particles in a Liquid. **STOKES' LAW.** Stokes (50) was the first to suggest the relation between the radius of a particle and its rate of fall in a liquid. He stated that the resistance offered by the

liquid to the fall of the particle varied with the radius of the sphere and not with the surface, according to the formula:

$$V = \frac{2 (d_p - d) g r^2}{9 \eta} \quad (4)$$

where V is the velocity of fall in centimeters per second, g the acceleration due to gravity, d_p the density of the particle, d the density of the liquid, r the radius of the particle in centimeters and η the absolute viscosity of the liquid. It is obvious that the velocity of fall of particles with the same density, in a given liquid, will increase with the square of the radius.

ASSUMPTIONS IN STOKES' LAW. There are several fundamental assumptions upon which the validity of Stokes' formula is based. These are:

1. The particles must be large in comparison to liquid molecules so that Brownian movement will not affect the fall.
2. The extent of the liquid must be great in comparison with the size of the particles. The fall of the particle must not be affected by the proximity of the wall of the vessel or of adjacent particles. Köhn (28) has suggested that particles closer than 0.1 mm. to the wall of the vessel encounter appreciable resistance to fall. Odén (38) has shown that particles fall independently of each other in suspensions of 1 per cent or less of solid particles. Concentrations of 2 per cent and 5 per cent have been used by Robinson (45) and Wiegner, respectively, for sedimentation analyses.
3. Particles must be rigid and smooth. This requirement is difficult to fulfill with soil particles. It is highly probable that the particles are not completely smooth over their entire surfaces. It is fairly well established that the particles are not spherical but are irregularly shaped with a large number of plate-shaped particles present in the clay fractions. Since variously shaped particles fall with different velocities, the term "equivalent or effective radius" is used to overcome this difficulty in Stokes' law. "Equivalent or effective radius" is defined as the radius of a sphere of the same material which would fall with the same velocity as the particle in question.
4. There must be no slipping between the particle and the liquid. This requirement is easily fulfilled in the case of soils because of the water hull around the particles.
5. The velocity of fall must not exceed a certain critical value so that the viscosity of the liquid remains the only resistance to the

fall of the particle. Particles larger than silt cannot be separated accurately by the use of Stokes' law.

LIMITATION OF STOKES' LAW. In addition to the effect of the size and shape of the particles upon the applicability of Stokes' law to mechanical analyses, there are certain experimental limitations that must be considered in the use of this principle.

Since the rate of fall varies inversely with the viscosity of the medium, it is necessary to maintain a known constant temperature during the analysis. For example, the velocity of fall at 30° C. is about 12 per cent faster than at 25° C.

A constant temperature helps also to prevent convection currents that might arise as a result of differences in temperature near the walls of the vessel and within the suspension. Any such currents would prevent the uniform settling of the particles. Convection currents may also be set up when the soil suspension is stirred. Currents due to stirring are more difficult to eliminate than those caused by temperature variations.

The density of the soil particle is another factor that affects the accuracy of Stokes' law. Density depends upon the mineralogical and chemical constitution of the particles as well as upon their degree of hydration. It is questionable whether the actual density of a particle as it falls in water can be accurately determined. The pycnometer method is most commonly used for measuring density. The samples are dried at 110° C. to expel the adsorbed water hull. It is obvious that a hydrated particle, falling in a suspension, should have a lower density than one that is completely dehydrated. Attempts have been made to investigate the hydration effect by using heavy aqueous liquids as a means of determining density. Marshall (31) has used solutions of potassium mercuric iodide of different densities. Flocculation of the clay particles takes place, and it is easy to observe if the liquid or the soil particle has the higher density. His results are shown in Table 6, along with other values for the specific gravity of different soil constituents.

It is readily seen from these data that a specific-gravity determination should be made for each soil in order to obtain the best results. The generally accepted value of 2.65 represents an average figure that is sufficiently exact for the majority of mechanical analyses. Accurate studies, however, require a more exact value. It is essential to note that the density of clay particles decreases with particle size when determined in aqueous suspension. This is due to the higher hydration of the smaller particles.

TABLE 6

THE DENSITY OF VARIOUS SOIL CONSTITUENTS

Orthoclase.....	2.50 to 2.60	Putnam clay.....	2.52 to 2.78
Mica.....	2.80 to 3.20	Cecil clay.....	3.35
Quartz.....	2.50 to 2.80	Rothamsted clay.....	2.74
Limonite.....	3.40 to 4.00	Putnam clay *	
Hematite.....	5.10 to 5.20	1-2 μ	2.56
Fe(OH) ₃	3.73	0.5-1 μ	2.50
Kaolin.....	2.50	0.2-0.5 μ	2.42
Humus.....	1.37	0.1-0.2 μ	2.39

* Determined with K₂HgI₃ solutions (31).

In light of the effects of particle size, shape and density upon the applicability of Stokes' law to mechanical analysis, it should be remembered that any particle which is separated upon the basis of its settling velocity does not necessarily have the exact calculated size. Its effective or equivalent radius corresponds to a given size grouping, in that all particles within such a group have the same velocities of fall. For this reason many investigators have followed the suggestion of Robinson (46) to define particle sizes in terms of their settling velocities rather than by their equivalent radii or diameters.

Although these limitations to Stokes' law may become serious in refined analyses of the size distribution of various particles in the soil, for most purposes the results will not warrant extra precaution in technique provided that extreme temperature fluctuations are prevented and care is exercised in the stirring of the suspension.

Analyses Based upon Complete Separation of Particles. SIEVE METHOD. Methods for making a mechanical analysis of soils can be divided into two distinct classes, namely, (1) those that achieve a complete separation of the various particle sizes into different groups and (2) those that express the mechanical composition as a continuous function of particle size.

Sieving was one of the first methods to be used for separating the coarser fractions of soils into groups of various sizes. The fact that the sieving technique is simple, rapid and inexpensive has made it a popular method of analysis. It is clear that if all the soil particles were round, sieves could be made which would permit accurate separations of coarse particles. It should be obvious also that the smallest cross-sectional area of irregularly shaped particles determines the size of the material which passes through the screen, and that an average diameter for such a group of particles may not be very exact. Moreover, the size of the opening in a given sieve increases with time as a result of

wear, a fact which must be taken into consideration in the constant use of sieves.

In spite of these possible objections, the use of the sieve gives satisfactory results as long as there is no significant change in the size of the opening. The great limitation of sieving is the fact that it can be used only for the coarser fractions. It is difficult to sieve particles smaller than 0.1 mm. in diameter, although Olmstead and his associates (40) have shown that a soil suspension can be washed through a 300-mesh screen which permits particles smaller than 0.05 mm. to pass through. This size limit was convenient in classes of the old sizes used by the Bureau of Chemistry and Soils, since it represented the upper limit of silt. If the size classes of the International Society of Soil Science are employed, however, the upper limit of 0.02 mm. cannot be achieved by sieving. Nevertheless, it is highly probable that most techniques for mechanical analyses will continue to employ sieves to separate particles larger than 0.05 to 0.1 mm. The fractions removed by sieving can be included in the summation percentage curves determined by other methods, to give a complete curve for the entire soil sample.

SEDIMENTATION METHODS. Sedimentation of particles and subsequent decantation of the material remaining in suspension have long been used to effect a mechanical analysis of the soil. The centrifuge may be employed to increase the rate of settling of the finer particles. The main principle of these techniques is to allow the soil suspension to settle, either by gravity or by centrifugation, until all particles larger than a given diameter have settled to the bottom. The material remaining in suspension is immediately decanted or siphoned off. Since some particles smaller than those in question have also settled to the bottom, it is necessary to repeat this operation until the supernatant liquid above the settled material is clear and free from fine particles. The time of settling may be calculated from Stokes' law or may be checked by measuring the size of the particles with a microscope. Perhaps the largest error introduced in the analysis by this technique is the carrying out of settled material during the decantation or siphoning processes.

The use of the Atterberg sedimentation cylinder or Schlammapparat (Figure 7) has been one of the classical methods in the development of sedimentation techniques for mechanical analyses (2).^{*} It consists

^{*}It is not the purpose of this book to present a complete historical review of all the methods of mechanical analyses. Such a discussion may be found in the book by W. C. Krumbein and F. J. Pettijohn, *Manual of Sedimentary Petrology*, Appleton-Century, 1938. The author attempts to call attention to the significant developments in technique as he sees them.

essentially of a graduated cylinder with a side outlet for removing the suspension above the sedimented material. This apparatus has been used rather widely in Europe. Even though the side outlet aims to remove only the suspended material above a given depth, considerable mixing of layers takes place as a result of turbulence within the cylinder during the removal of the soil suspension. These turbulent effects have been studied by Köhn (28).

The Osborne beaker method was the earliest sedimentation technique used in the United States. A series of different-sized beakers was employed. The time of settling was determined by checking the size of the particles with the microscope. This method was the basis for the technique used for a long time by the United States Bureau of Soils, which depended upon gravity sedimentation to separate the sands from the silt and clay and centrifugal sedimentation to remove the silt from the clay. The time of settling and centrifuging was determined by microscopic examination of the material in suspension above the sedimented material. The process of redispersion, sedimentation and decantation was continued until all the clay had been separated from the silt and the silt from the sands. The sands were subdivided by sieving. Although this technique was laborious, especially with soils containing much clay, a careful analyst could obtain very good results.

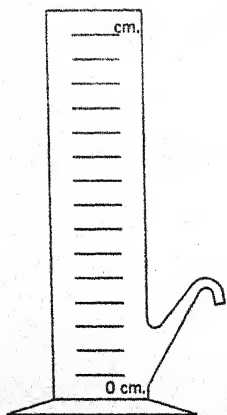


FIG. 7. The Atterberg Schlammapparat.

Centrifugation. In applying Stokes' law to the velocity of fall of particles during centrifugation it is necessary to remember that the centrifugal force exerted upon a given particle depends upon the distance of that particle from the axis of rotation. This is illustrated in the following equations:

$$\text{Centrifugal force} = \frac{(\text{linear velocity})^2}{\text{radius of rotation}} \text{ or } \frac{V^2}{R} \quad (5)$$

but $V = \text{radius of rotation} \times \text{angular velocity}$, or $R\omega$. Therefore,

$$\text{Centrifugal force} = R\omega^2 \quad (6)$$

During centrifugation, the value of R for a given particle increases as that particle is sedimented. Consequently, it is not sufficient to substitute $R\omega^2$ for g in Stokes' law without taking into consideration the change in R as a function of time. When these facts are considered, Stokes' law may be written as:

$$\frac{dR}{dt} = \frac{2}{9} \frac{R\omega^2 (d_p - d)r^2}{\eta} \quad (7)$$

Integrating this expression between the limits of t_0 and t and the corresponding limits, R_1 and R_2 :

$$\int_0^t dt = \frac{1}{\frac{2}{9} \frac{\omega^2 (d_p - d)r^2}{\eta}} \int_{R_1}^{R_2} \frac{dR}{R} \quad (8)$$

The time required to centrifuge a particle from the surface (R_1) of a suspension in a centrifuge tube to a given depth (R_2) becomes:

$$t = \frac{9\eta \ln \frac{R_2}{R_1}}{2\omega^2 r^2 (d_p - d)}$$

Since $\omega^2 = (2\pi N)^2$, where N = number of revolutions per second, then

$$t = \frac{9\eta \cdot 2.303 \log \frac{R_2}{R_1}}{2 \cdot 39.44 N^2 r^2 (d_p - d)} = \frac{\eta \log \frac{R_2}{R_1}}{3.81 N^2 r^2 (d_p - d)} \quad (9)$$

Steele and Bradfield (49) have used this formula to advantage with the pipette method, as will be discussed later.

The supercentrifuge, which follows the principle of the cream separator, is used frequently to separate various clay fractions. The bowl is a long cylinder through which the suspension is passed during centrifugation. Speeds of rotation up to 40,000 or more revolutions per minute can be obtained. Colloidal particles can be easily thrown out on the inner walls of the cylinder by such a high centrifugal force. The size of particles that will be deposited depends upon the speed of rotation and the rate at which the suspension passes through the bowl. Ayers (3) has made a very thorough study of the problem of separation with the supercentrifuge. According to him the extent to which the velocity of a settling particle is increased by the supercentrifuge is determined by two factors: first, by the number of times the force of gravity (f_0) is increased and, second, by a concentration factor (k_1). The value k_1 is expressed as parts per hundred by weight of suspended material in the suspension. In other words, if the concentration of the suspension is .1 per cent, the value k_1 is equal to 1.0 (not 0.01). It must be introduced because it enters into the calculation of both the centrifugal force on all the particles at any point and the

osmotic pressure exerted by these particles at the same point. Ayers has pointed out that the only considerable force opposing settling in such a separation is the osmotic pressure. This must be subtracted from the centrifugal force to give the actual or resultant forces on the particles.

The approximate equation for velocity given by him is:

$$V = \frac{2r^2 k_1 f_0 g (d_1 - d_2)}{9\eta} \quad (10)$$

where V is the velocity of particles toward periphery of the bowl (cm./sec.); r the radius of particles (cm.); k_1 the parts per hundred of suspended material; f_0 a factor multiplying gravity; g the acceleration due to gravity (cm./sec.²); d_1 the density of particles (gm./cm.³); d_2 the density of medium (gm./cm.³); and η the viscosity of suspension (poises). It should be borne in mind that this equation is only approximate, but it is believed to be as accurate as Stokes' law in gravity subsidence.

The equation for velocity can be transformed into an equation for time, that is, the time necessary for particles of a definite size to settle out, or, in other words, travel to the periphery of the bowl. Particles below this size will pass out of the bowl at the upper outlets.

There is a cylinder of space in the center of the bowl when the centrifuge is in operation. The diameter of this cylinder is equal to that of the circle of openings in the top of the separator bowl. The depth of liquid, or horizontal distance a particle must travel after entering the bowl before it can be considered settled, is equal to the radius of the bowl minus the radius of the cylinder of space. The time for a complete refilling of the bowl can be calculated as follows:

$$\text{Time} = \frac{9\eta S}{2r^2 k_1 f_0 g (d_1 - d_2)} \quad (11)$$

where S is the depth of liquid in bowl.

For accurate separations, the value k_1 must be determined prior to each time the material is passed through the supercentrifuge. If the concentration of the suspension is kept below 0.4 per cent, the viscosity does not change sufficiently to merit its redetermination.

It is necessary to redisperse the material collecting on the bowl (a celluloid sleeve inserted in the bowl to receive the deposited particles is most convenient) and to centrifuge until the liquid coming out of the bowl is practically free from suspended particles. For example, a

sample of Putnam clay had to be centrifuged 26 times to remove completely particles from 100 to 500 $m\mu$ in diameter (54). The properties of the various fractions removed by this technique were almost identical with those separated from the sample of Putnam clay by means of the two-layer method. (See discussion on page 13.)

It is exceedingly difficult to perfect a quantitative separation with the supercentrifuge, even though qualitative separations into various fractions are easily possible. It is highly probable that the use of the supercentrifuge in clay separations will be limited to removing desired quantities of a given fraction. At the present time the supercentrifuge is used primarily for separating the colloidal fraction (smaller than 100 $m\mu$) from clays.

Most sedimentation methods for the complete separation of soil particles have the advantages of simplicity, inexpensiveness and easy adaptability to routine analysis. They have the serious disadvantage, however, of magnifying a small error in technique, owing to the fact that the decantation process must be repeated, and this error is multiplied by the number of repetitions. It is obvious that the currents which are set up during shaking and decantation or siphoning are likely to cause significant inaccuracies in the analysis, since it is difficult to remove a given column of suspension without some intermixing of layers. Moreover, even with the centrifuging, ordinary methods of analysis which depend upon complete separation of the different fractions cannot be easily extended to particles with diameters below 0.002 to 0.001 mm.

Marshall (31) has recognized the importance of developing a technique which is applicable to particle separations below 0.002 mm. He has visualized that the ideal mechanical analysis will permit all particles of different sizes to move from the same plane at a given time. To achieve this, a thin layer of clay suspension is floated upon a denser column of sugar or glycerine solution in a centrifuge tube. This system is centrifuged at a speed sufficient to cause all particles of a given diameter to settle through the denser solution beneath the suspension. Smaller particles will remain in the supernatant liquid. In this way a complete separation of a given fraction is accomplished in one centrifuging. This technique is known as the **two-layer method** in mechanical analysis. A schematic representation of the layers in the tube is shown in

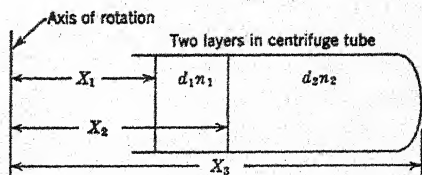


FIG. 8. Schematic representation of the two-layer method of Marshall (31).

Figure 8. The following formula is used to calculate the speed of centrifugation for various particles:

$$t = \frac{1}{2} t_1 + t_2 = \frac{1}{3.81 r^2 N^2} \left[\frac{\eta_1 \log \left(\frac{x_2}{x_1} \right)}{2(D - d_1)} + \frac{\eta_2 \log \left(\frac{x_3}{x_2} \right)}{D - d_2} \right] \quad (12)$$

where t is the average time for a spherical particle of radius, r , and density, D , to reach the bottom of tube, t_1 and t_2 are the times necessary to settle through upper and lower layers, respectively, η_1 and η_2 are the viscosities of upper and lower layers, respectively, d_1 and d_2 are the densities of upper and lower layers, respectively, x_1 , x_2 and x_3 are the distances from the axis of rotation to the top of the upper layer, to the boundary between the two layers and to the bottom of the tube, respectively, and N is the number of revolutions per second.

Marshall has obtained very reliable results with separations of particle sizes as small as 50 $m\mu$. Best results have been obtained with a 1 : 4 glycerine-water mixture for particles larger than 100 $m\mu$. This technique requires a centrifuge that can be driven from 1500 to 4000 revolutions per minute without developing enough heat to cause a change in viscosity of the two layers. Moreover, the acceleration and retardation of the centrifuge must be controlled sufficiently to prevent the disturbance of a clear-cut boundary between the two liquids. Most centrifuges that are used for ordinary mechanical analyses are not well adapted to the two-layer method, except in the coarser clay region which does not require an excessive speed of centrifugation. Dilute clay suspensions must be used for the greatest accuracy; about a 0.2 per cent suspension is recommended. This method has many possibilities for accurate separations.

Whiteside and Marshall (52) have adapted the two-layer principle to the supercentrifuge and have been able to make particle separations below 10 $m\mu$ in diameter with good results. The adaptation of this technique to the supercentrifuge offers many possibilities for making separations at the lower end of the colloidal range and for studying the chemical and physical properties of these fine colloidal fractions.

ELUTRIATION METHODS. In sedimentation methods of mechanical analysis, particles of various sizes are separated on the basis of their settling velocities in still water. Elutriation methods separate different soil fractions by "washing out" particles in a rising current of water. Water is allowed to flow through a vertical cylinder from the bottom to the top with a given velocity. If a soil suspension is placed in this elutriation tube, all particles that have settling velocities smaller than

the upward velocity of the water will be carried out of the tube. Separations into fractions of different sizes can be achieved by changing the velocity of the current through a single tube or by passing the elutriated material through a series of tubes of different diameters. The diameter of a given tube will determine the upward velocity of water in that tube if the flow of water into the apparatus is constant.

Although numerous types of elutriators have been used in the mechanical analysis of soils, most of the various methods are based upon the single-tube elutriator as developed by Schöne in 1867, and the multiple-tube type as exemplified in the well-known Kopecky apparatus.

Schöne (48) was among the first to recognize that the various shapes, sizes and densities of particles did not permit a rigid separation of particles having the exact size as calculated by Stokes' law. The term **equivalent hydraulic value** was suggested as giving a more accurate conception of the size of particles that were being separated. Schöne applied this concept in the development of an elutriation technique for mechanical analysis of soils.

The apparatus (Figure 9) consists of a cone-shaped vessel about 20 inches tall with an inlet tube at the bottom and an opening at the top.



FIG. 9. The Schöne elutriator.

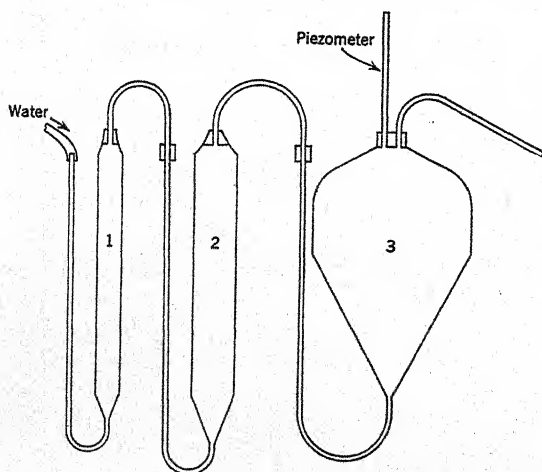


FIG. 10. The Kopecky elutriator.

The glass tube at the top is so constructed as to provide not only an outlet for the elutriated material but also a pressure or piezometer tube to check the velocity of the current through the elutriator. The outlet opening is made just small enough to cause a back pressure which is reflected by the length of the column of suspension in the vertical part of the tube.

To make a mechanical analysis, a properly dispersed sample of soil is placed in the apparatus and a current of water is passed through with a velocity of flow just great enough to remove the smallest-sized particle that

is desired in the analysis. When the water comes out of the apparatus clear, the velocity is increased to remove the particles of next largest size. This process is continued until the desired analysis is complete.

The Kopecky (29) elutriator (Figure 10) consists of three vessels of different diameters, which are placed side by side. The elutriated material from the first vessel enters the bottom of the second; the elutriated material from the second vessel enters the bottom of the third. A piezometer tube is placed on the top of the third tube. The diameters of the original Kopecky tubes were 3.0, 5.6 and 17.8 cm., respectively. The narrow diameter of the first vessel causes a high upward velocity of the rising current of water which carries all but the coarsest particles into the second tube. The increased diameter in the second tube causes a decrease in the upward velocity of water, and a given group of particles remain behind in this tube when elutriation is complete. A similar effect takes place in the third vessel. When elutriation is completed, as evidenced by the absence of particles in the water leaving the last tube, four distinct separations are achieved. The size limits of these four groups will depend upon the rate at which water enters the first tube and the diameter of each of the tubes. The shape of the last tube is the most serious limiting factor in the Kopecky apparatus. The length of that portion of the tube having vertical side walls is not sufficient to overcome convection currents which occur during the entry of water into the bottom of the vessel.

Elutriation methods have the advantage of simplicity of operation, once the exact velocity of flow of water through the different tubes has been ascertained. Little attention is required after the separation of particles has been started. Moreover, the current serves to separate flocules and loosely held aggregates. As will be discussed in Chapter V, elutriation offers considerable possibilities in determining the stable aggregates in soils.

There are several distinct disadvantages in the use of elutriation techniques for mechanical analyses. It is difficult to regulate the flow of water through the various tubes and to make sure that the velocity in a given tube is the one desired. Serious convection currents are set up in all the tubes. As the water enters the narrow opening at the bottom there is a tendency for a greater velocity of flow to take place in the center of the tube. This effect is dissipated if the tube is long enough to permit uniform flow at the top. Moreover, particles fall back on the cone-shaped part of the elutriator and are not completely washed by the rising current. Variations in the shape of the

tubes affect the results. These effects are more serious in the wider vessels where the velocity of the water is reduced. Longer tubes help to obviate these difficulties.

Perhaps the most serious disadvantage of elutriation is the narrow range of operation as far as particle sizes are concerned. A particle about 0.02 mm. in diameter is the smallest that can be separated because the flow of water is too slow for smaller fractionations. If the flow of water in the last vessel is slow enough to remove 0.02 mm. particles, the first tube must be very narrow to permit separations larger than 0.2 to 0.5 mm. In light of these difficulties, it is doubtful if elutriation techniques will ever find wide application in mechanical analysis of soils, with the possible exception of determinations of the stability of soil granules.

Analyses Based upon the Size Distribution of Soil Particles.
THEORETICAL CONSIDERATIONS. The methods of mechanical analysis which have been discussed in the preceding sections permit the separation of the soil particles into a number of distinct fractions between certain size limits. It is obvious that, if the number of separated fractions could be sufficiently increased, one would obtain a picture of the percentage distribution of the particles of different sizes in the sample. Such a complete analysis by any of the aforementioned methods, with the possible exception of the two-layer technique of Marshall, would be too laborious and inaccurate to be practical.

In 1915, Odén (36) developed a theory of the settling of particles in suspension, as based upon Stokes' law, and gave it a mathematical interpretation. Later, Fisher and Odén (14) presented a more general analysis of the theory, which has served as the basis for most of the present methods of mechanical analysis. A simple interpretation of their analysis is given in the following paragraphs.

Let us consider a uniform suspension of soil particles in which all particles are settling independently of each other. At any time t and at a given depth h , those particles with settling velocities greater than h/t cm. per sec. will have settled below the depth. In addition, particles with settling velocities smaller than h/t cm. per sec., but falling through a shorter distance, will have settled below h . If the pan of a balance is placed at the depth h , the total weight of sedimented particles at the end of time t will be equal to the weight of that fraction with a settling velocity of or greater than h/t cm. per sec., that has settled out completely, plus a portion of the material with a settling velocity less than h/t . The actual rate of increase in weight dp/dt at time t is due to this latter group of particles, since they have been falling at constant rate since the beginning of the experiment. The

actual weight of these smaller particles on the pan of the balance is $t dp/dt$.

Perhaps the concept of the settling of particles in suspension will be much clearer if a hypothetical case is discussed. Visualize a suspension containing only two different sizes of particles. The height of the suspension above the pan of a balance is 10 cm. The suspension contains 2 gm. of particles that will settle out completely in 20 minutes and 2 gm. of particles that will sediment in 40 minutes. The settling velocities, therefore, are $1/120$ and $1/240$ cm. per sec. (h/t), respectively. At the end of 20 minutes all the faster-sedimenting particles will settle on the pan. During sedimentation they will increase the weight at the rate of $1/600$ gm. per sec., since $1/1200$ of this fraction will settle out every second. During this same time, the smaller particles will settle at the constant rate of $1/240$ cm. per sec. Since $1/2400$ of the 2 gm. will sediment every second, these particles will increase the weight at the rate of $1/1200$ gm. per sec. (dp/dt). At the end of the 20 minutes 1 gm. of this material will

settle out $\left(t \frac{dp}{dt} = 1200 \times \frac{1}{1200} \right)$. Therefore, at the end of time t (20 minutes) the rate of increase in weight will be due to the 1 gm. of these smaller particles that will continue to settle out for 20 more minutes at a constant rate of $1/1200$ of a gram per second. It is obvious that the same reasoning will apply to an indefinite number of fractions with settling velocities of less than h/t cm. per second.

Instead of visualizing the pan of a balance at a given depth, onto which the falling particles settle out at different rates to cause a continuous increase in weight, one might consider the changes in concentration that take place in a thin plane of the suspension at this same depth. At the beginning of sedimentation, every plane in the suspension has the same number of particles of different sizes. As sedimentation proceeds, however, particles with settling velocities greater than h/t cm. per sec. will settle from the top through the plane at depth h . Smaller particles will settle out of this plane, but will be replaced by similar particles that have fallen from above into this thin layer. As the time of settling increases, it is seen that there is a progressive decrease in concentration or density of the layer at this given depth.

These changes in concentration may be calculated according to the following reasoning: Let W equal the weight of particles in a given volume, V , of suspension, D_p the density of the particles, D_s the density of the suspension, v the settling velocity of the particles, t the

time of settling, h the depth of settling and X the fraction or percentage by weight of particles having a velocity less than $v = h/t$.

At time t_0 in a completely dispersed, uniform suspension, every cubic centimeter of the suspension will contain W/VD_p cc. or W/V gm. of solid particles and $(1 - W/VD_p)$ cc. or gm. of water. The specific gravity of the suspension at the beginning will be:

$$D_{s0} = 1 + \frac{W}{VD_p} (D_p - 1) \quad (13)$$

If X equals the fraction of particles having settling velocities less than h/t , then at time t and depth h there will be left in suspension XW/V gm. of particles per cc. immediately above h , because the faster-moving particles will have settled below this depth. The particles included in X will be in the same concentration as they were at the start. At time t , therefore, the specific gravity of the suspension in a thin layer above h will be:

$$D_s = 1 + \frac{XW}{VD_p} (D_p - 1) \quad (14)$$

or

$$D_s = 1 + kX \quad (15)$$

This equation is fundamental to the several methods that are based upon the Odén theory.

FUNDAMENTAL REQUIREMENTS. There are several fundamental requirements of methods based upon the change in concentration of a suspension at a given depth with time in order to permit an accurate measurement of this change. First, it is necessary to have complete dispersion of the soil particles and to prevent any coagulation effects during settling. It is obvious that an incomplete dispersion will not permit a true measurement of the quantity of a given fraction. Unlike methods based upon repeated fractionations where dispersion may be effected as the separations proceed, the principle of determining changes in concentration requires that each particle fall independently of the others.

Second, the concentration of the suspension must be dilute enough so that the particles do not interfere with each other during their fall through the liquid. Moreover, the density of the suspension should never vary greatly from that of the liquid. Odén advocated the use of a 1 per cent suspension; Robinson used a 2 per cent; and Wiegner carried out most of his investigations in a concentration of 5 per cent. Not only must the suspension be dilute but the sedimentation vessel

also must have a diameter large enough to prevent errors arising from the attraction between the particles and the wall.

Third, the suspension must be thoroughly stirred so that there is a uniform distribution of particles throughout at the beginning time, t_0 . All methods presuppose that every particle settles at a uniform rate from all portions of the suspension. This fact will be much clearer after the discussion of the pipette method on page 53.

Fourth, measurements should be made at a constant temperature so that no convection currents occur during sedimentation. Differences in temperature between the wall of the vessel and the center of the suspension may produce currents that will prevent the uniform settling of the finer particles. It is also essential to maintain a constant temperature in order not to change the viscosity of the dispersion medium. Although these four requirements may seriously affect the results of size-distribution analyses, if improper techniques are employed they are so easily met under good experimental conditions that they are not a handicap to routine determinations.

GRAPHICAL SOLUTION OF ACCUMULATION AND DISTRIBUTION CURVES. Inasmuch as Odén measured the amount of material that sedimented at different time intervals, he called the curve, which was obtained by plotting time on the abscissa and the weight of deposited particles on the ordinate, an **accumulation curve**. In other words, this curve represents the total accumulated material at various time intervals (Figure 11c). From this curve the **size-frequency distribution curve** is obtained by graphical analysis. The **size-frequency distribution curve** depicts the relative amount or distribution of particles as a function of their size. The ordinate is a function of the radius $F(r)$, and the abscissa is the radius or diameter of the particles. In other words, the ordinate is a relative expression of the amount of particles in each size range. The percentage of particles between any two radii, r_1 and r_2 , is obtained simply from the surface or area encompassed by the axis of the abscissa, the corresponding ordinates and the curve. This is represented by the area $ABCD$ in Figure 11a. It should be obvious that the peaks in the distribution curves represent the size of particles that are present in greatest numbers. If one wishes to visualize this distribution mathematically instead of graphically, the relative amount of particles between r_1 and r_2 will be

expressed by the integral $\int_{r_1}^{r_2} F(r)dr$. In order to condense these

graphs into a more compact form, it is often convenient to plot the logarithm of the radius on the abscissa instead of the radius.

It is not the purpose of this book to delve into the complicated mathematical solution of accumulation curves to obtain the data from which the size-frequency distribution curves are plotted. Such a treatise may be found in the paper by Fisher and Odén (14). Since most of the distribution curves, however, are obtained by a graphical analysis of the accumulation curves, this type of solution will be presented somewhat in detail.

It has been pointed out on page 43 that the rate of change in concentration of a thin layer of soil suspension at a given depth h at any

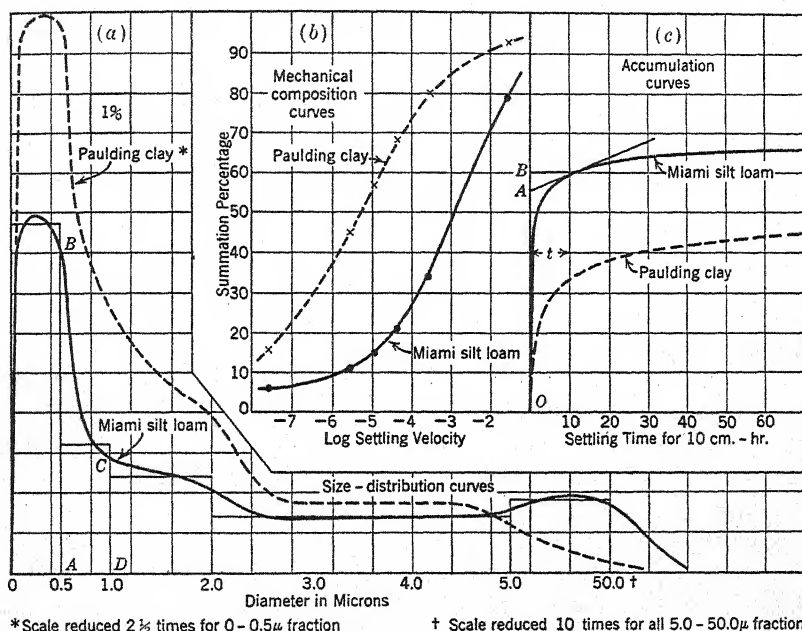


FIG. 11. Accumulation (c), size-distribution (a), and mechanical composition (b) curves of Paulding clay and Miami silt loam.

time t will be due to those particles that have settling velocities less than h/t cm. per sec. If the pan of a balance is placed at depth h , then the rate of change in weight will be due to the settling of the smaller particles. Since this rate of change is characterized by the slope of the accumulation curve at time t , a tangent to the curve at this point will give the settling rate (gm/t) of this group of particles. Inasmuch as they were sedimenting at this constant velocity from the beginning, the extrapolation of this tangent so as to intercept the y -axis (weight or percentage of particles) at t_0 will permit the calcu-

lation of the amount of these smaller particles that have settled out in time t . The following data can be obtained from the accumulation curve (Figure 11c) by this process:

1. The total weight or percentage, W , at time t will equal the length of the ordinate from this point of tangency, or OB .
2. The weight or percentage of the smaller particles that have settled out will equal the difference between the ordinate corresponding to W and the ordinate as represented by the cutting of the y -axis with the tangent; that is, $OB - OA = AB$.
3. The weight or percentage of particles having radii equal to or greater than that corresponding to the settling velocity of h/t cm. per sec. will be equal to the length of the ordinate as represented by the cutting of the y -axis with the tangent, or OA .

By repeating this process for various values of t , which are chosen so as to correspond to the desired radii, it is possible to obtain the amounts of the various sizes of particles present.

A clearer concept of this type of graphical analysis may be obtained from the different graphs in Figure 12. Figure 12a shows the settling curves for suspensions containing only one size of uniform particles. Line a_1 represents the cumulative settling of coarse particles which sediment at the rate of 4 gm. per 20 min.; line a_4 shows the settling of fine particles which sediment at the rate of 1 gm. per 20 min. If each suspension contains 4 gm. of particles at the outset, then a_1 will settle out completely in 20 minutes. It would take a_4 80 minutes to sediment.

Consider now the experimental setup in Figure 12b. This suspension contains 2 gm. of particles (a_1) with a settling time of 20 minutes (1/2 cm. per min.), and 4 gm. of material (a_2) with a settling time of 40 minutes (1/4 cm. per min.). At the end of the first 20 minutes all of a_1 will have sedimented and one-half of a_2 , which will make a total of 4 gm., as shown by the point, P_1 . At the end of this time, the remainder of a_2 will continue to settle out at the uniform rate of 1/4 cm. per min. and will completely sediment at the end of 40 minutes, as indicated by the point, P_2 . If the uniform rate of settling line for a_2 , between the points P_1 and P_2 , is extrapolated so as to intercept the y -axis at A , it is seen that OA and AB both equal 2 gm. of material. This graphical process is nothing more than drawing a tangent to the curve at the point P_1 . The total weight of sedimented particles at P_1 is equal to OB , or 4 gm.; the weight of the smaller particles (a_2) that have settled out is equal to AB ($OB - OA$), or 2 gm.; and the weight of the larger particles (a_1) is equal to OA , or 2 gm.

In Figure 12c, four different separates are mixed together. At the end of the first 20 minutes all of a_1 will settle out and one gram each of the other separates. A tangent to the point P_1 shows that one gram of the coarsest fractions has sedimented. At the end of the next 20 minutes, the remaining one gram of a_2 will settle out plus another gram each of a_3 and a_4 . A total of 7 gm. will have sedimented at this time. Of this amount, 3 gm. will be the coarse fractions a_1 and a_2 , and 4 gm. will be due to the finer fractions. At the end of 60 minutes, a tangent at the point, P_3 , intercepts the y-axis at C and shows that

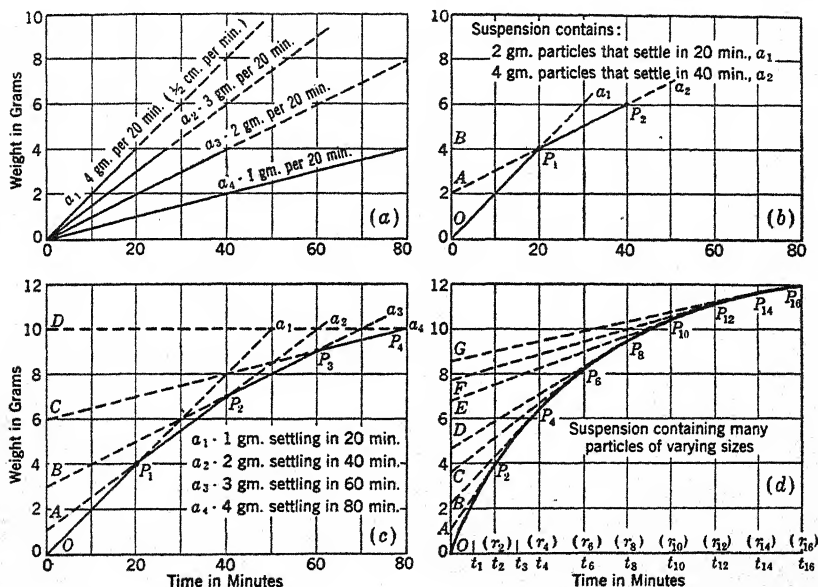


FIG. 12. Graphical analysis of accumulation curves (developed by Byron Shaw).

the suspension contains 4 gm. (DC) of a_4 material; the tangent at P_2 indicates that there are 3 gm. (BC) of a_3 particles.

Figure 12d shows how a series of tangents to a smooth accumulation curve obtained from a polydispersed suspension may be used to determine the amount of material deposited at t_1, t_2 , etc., which correspond to radii, r_1, r_2 , etc., respectively. OA is the weight of material larger than r_2 ; AB is the amount between r_2 and r_4 ; the other fractions are obtained in like manner.

Once the weight or percentage of particles between the desired size limits is obtained from the accumulation curve, they are plotted in a histogram to get the size-frequency distribution curve. A small area is chosen to represent a given weight or percentage, and the total

amount of particles within each size group is shown on the graph as an area. This is illustrated in Figure 11a. A smooth curve is then drawn across the tops of the histogram. The completed curve shows the frequency distribution of the various-sized fractions.

G. W. Robinson (45, 46) has suggested the use of mechanical composition curves to express the mechanical makeup of the soil. These curves are obtained by plotting the summation percentages against the logarithm of the settling velocities. (Settling velocity is used instead of particle diameter because particle size as obtained from Stokes' law implies only an equivalent or effective diameter due to variations in shape.) Several of these curves are illustrated in Figure 11b. The slope of the mechanical composition curves is a measure of the frequency of the fraction conforming to the corresponding settling velocity. The most abundant fraction is given by the steepest portion of the curve if the fractions are chosen at equal logarithmic intervals. Although these curves are based upon the very sound concept of settling velocities, it is believed that size-frequency distribution curves can be used to greater advantage from the standpoint of ready interpretation.

METHODS USED FOR DETERMINING DISTRIBUTION OF PARTICLES. Methods used to determine the size-frequency distribution of particles may be classified into six types, according to Odén (36, 37).

1. Changes in density of the suspension with time at a given depth. The pipette and hydrometer methods are classical examples of this principle.
2. Changes in density of the suspension with depth at a given time.
3. Changes in hydrostatic pressure of the suspension with time at a given depth. This principle is illustrated in the Wiegner method.
4. Changes in hydrostatic pressure of the suspension with depth at a given time.
5. Changes in weight of an immersed body in the suspension with time at a given depth. The plummet method of Olmstead is an example of this principle.
6. Changes in weight of sediment deposited from suspension with time at a given depth. The Odén automatic balance is the best-known example of this type.

Odén's Automatic Balance. Odén obtained data to support his theory of the settling of particles in dilute suspension by using an automatic, sedimentation balance. The main features of this appara-

tus are shown schematically in Figure 13. One pan of the balance, *A*, is suspended near the bottom of the sedimentation cylinder, *C*. The other pan, *B*, is counterpoised with *A*. As particles settle on *A* they cause it to lower; electrical contact is made at *E* which releases a small shot of known weight onto *B*. By knowing the number of shots and the time after settling began that each shot was released, a curve can be drawn to show the accumulated weight of sediment as a function of time. This ingenious device was later modified by Coutts, Crowther, Keen and Odén (11) to make it self-recording. Full details of the Odén-Keen balance may be found in the original article or in Keen's book (26).

The main advantages of the Odén-Keen balance are its automatic self-recording features. There are several serious disadvantages in its use which have prevented any wide employment of this technique. In the first place, the required equipment and experimental setup are beyond the reach of most research laboratories. In addition, there are

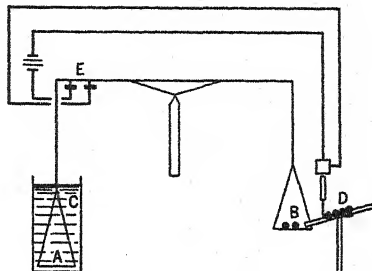


FIG. 13. Sketch of the fundamental principle of the Odén automatic balance (38).

several inherent errors in the technique that are difficult to surmount. As the particles settle out, a clear liquid is left immediately beneath the pan, owing to the fact that the falling particles have been intercepted by the pan. This is not true between the pan and the wall of the cylinder. Consequently, the difference in density between the suspension next to the wall and the liquid beneath the pan is in part responsible for currents that are formed throughout the suspension, which prevent the independent fall of the particles. Moreover, the currents that are set up during the stirring of the suspension along with convection currents cause a slight accumulation of the particles near the wall of the cylinder.

Wiegner Sedimentation Cylinder. In 1918 Wiegner (55) developed the method of measuring changes in the hydrostatic pressure of a suspension at a given depth with time. This method was based upon the fact that when two columns of liquid of different densities are placed in two separate parallel tubes that are joined to each other, the liquids will rise to levels which are inversely proportional to their densities. Wiegner's sedimentation cylinder consisted of a long glass cylinder to which was attached a parallel tube of smaller diameter but of equal length that served as a manometer (Figure 14). A stopcock

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was placed in the smaller tube to permit separate filling of each arm of the apparatus. The apparatus is used in the following manner:

The soil suspension is placed in the larger tube and water in the smaller. The suspension is shaken and the stopcock is opened. As soon as the stopcock is opened, the height of the liquid in the manometer tube reflects the hydrostatic pressure at the junction of the two tubes. Since the specific gravity of the suspension is greater than that of water, the meniscus in the measuring tube will be higher than that of the suspension. The difference in level between the two tubes will be proportional to the pressure resulting from the weight of the soil particles above the level of the side arm. As the particles fall below the point of juncture with the side arm they no longer contribute to the hydrostatic pressure at this junction. Consequently, the density of the suspension decreases with time at this point and the pressure registered by the manometer decreases. The decrease in the height of the water in the side arm may be plotted as a function of time, to obtain a continuous curve from which the frequency curve can be obtained. Gessner (16) modified the original Wiegner technique by adding an automatic, recording device which photographed the position of the water in the side arm with time.

Wiegner used the following formula in the calculation of his results:

$$\frac{H_t - h_t}{H_0 - h_0} = \frac{W_t}{W_0} \quad (16)$$

where, W_0 and W_t are the weights of soil in the suspension at times t_0 and t , respectively, H_0 and H_t are the heights of water in the manometer tube at times t_0 and t , respectively, and h_0 and h_t are the heights of suspension in the sedimentation tube at times t_0 and t , respectively. The difference in heights of the menisci at any time is proportional, therefore, to the amount of soil that is present in the suspension.

The Wiegner apparatus is very simple and inexpensive. It is somewhat cumbersome to handle, especially when it is being shaken to get uniform dispersion. Care must be exercised to prevent convection currents and to obtain accurately the changes in the level of the side arm. Numerous modifications have been suggested to increase the accuracy of the readings. The most noteworthy have been the introduction of an auxiliary side arm at the top of the sedimentation cylinder and the bending of both manometric tubes at an angle. These modifications make it possible to follow both menisci more accurately be-

cause the tubes can be placed adjacent to each other, and the angle causes a larger movement of the menisci as the hydrostatic pressure changes.

Pipette Method. The pipette method of mechanical analysis is the most widely accepted technique which makes use of the Odén theory. Its simplicity has been one of the dominant factors in its almost universal usage. Variations in this method were introduced at about the same time (1922-1923) by Robinson (45) in England, Krauss (30) in Germany and Jennings, Thomas and Gardner (21) in the United States. The fundamental principle of the pipette technique is to determine the density of a suspension at a given depth as a function of time. Variations in density are measured by taking out samples of a definite volume at the required depth and determining the dry matter contained therein. The concentration of the suspension at depth h will be equal to the sum of the partial concentrations of

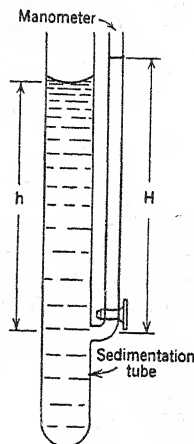


FIG. 14. The modified Wiegner apparatus.

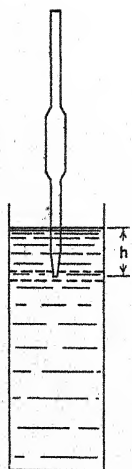


FIG. 15. The pipette method of mechanical analysis.

all fractions having settling velocities less than h/t cm. per sec. The ratio of the concentration at depth h after time t to the total concentration at the beginning of the experiment will give the proportion of material having velocities less than h/t . By determining the concentration for different values of h/t , a complete summation curve is obtained. This is illustrated in the following example:

Let us assume a soil suspension of 1000 cc. that contains 10 gm. of soil. At t_0 , each 25-cc. portion contains 0.25 gm. of particles. At time t_1 at a depth of 10 cm., 25 cc. of suspension are removed and evaporated to dryness and weighed. This volume of suspension contains 0.15 gm. of particles which have settling velocities less than $10/t_1$ cm. per sec., since all particles with greater velocities have settled from the top of the suspension past the 10 cm. depth. Therefore, there are $\frac{0.15}{0.25} \times 100$, or

60 per cent by weight, of particles in the original sample that have settling velocities less than $10/t_1$ cm. per sec. The suspension is shaken again and at time t_2 a second 25-cc. sample is removed from the 10-cm. depth. The second pipetteful contains 0.08 gm. of particles with velocities less than $10/t_2$, which are equal to

32 per cent, $\frac{0.08}{0.25} \times 100$, of the original soil sample. From these two samplings it is possible to make three separations:

Particles with settling velocities greater than $h/t_1 = 40$ per cent.

Particles with velocities less than $h/t_2 = 32$ per cent.

Particles with velocities between h/t_1 and $h/t_2 = 60 - 32 = 28$ per cent.

This procedure may be used to obtain any desired number of fractions. It is obvious that the same results would be obtained if the weight of particles in each pipetteful of suspension withdrawn is multiplied by the appropriate factor to give the total weight of these particles in the whole suspension.

Robinson employed an ordinary pipette in developing his particular technique; Krauss used a series of three pipettes with horizontal openings to remove the suspension from a given layer; Jennings, Thomas and Gardner employed a multiple pipette that was fixed in the suspension at the required depth. Olmstead and associates (40) studied the application of the pipette method to mechanical analysis and developed a technique with a single pipette that is now used by the soils laboratories of the United States Department of Agriculture. The use of a rack and pinion is almost essential to lower the pipette into the suspension at the required depth with the least disturbance of the liquid.

Steele and Bradfield (49) have applied the centrifuge to the pipette method with satisfactory results. They used 25 cc. of suspension in 100-cc. centrifuge tubes and removed a 5-cc. sample at a depth of 2 cm. R_1 and R_2 in equation 9 on page 37, therefore, were 22 and 24 cm., respectively. Using equation 9 for calculating the time of centrifuging and equation 4 for gravity sedimentation, they employed the pipette and accomplished separations from 5μ down to $62.5 \text{ m}\mu$. The various depths and times of sampling are given in Table 7. Time was taken from the moment the motor was turned on until it was turned off. No correction was applied for differences in the time of acceleration and retardation.

There are several factors that affect the values obtained by the pipette method. It is assumed theoretically that the pipette removes a thin layer of liquid at the required depth. Actually, however, a sphere of liquid is removed whose volume equals that of the pipette; the center of the sphere is at depth h . Köhn (28), using a solution of lead iodide which was strongly illuminated from the side, photographed the streamlines as the liquid entered the pipette and showed a sphere of the suspension was removed (Figure 16). Since a part of the sphere was on each side of the theoretical thin layer, the errors involved were considered to be compensated to such an extent as to be of little practi-

TABLE 7

DEPTH AND TIME OF SAMPLING SOIL SUSPENSIONS IN MECHANICAL ANALYSES AT 25° C. DENSITY OF PARTICLES = 2.65 (STEELE AND BRADFIELD (49))

Depth of sampling, cm.	Time		Size of particles, equivalent diameter in microns
	By gravity	By centrifuge 2200 R.P.M. $R_1 = 22$ cm. $R_2 = 24$ cm.	
	hr. min. sec.	hr. min. sec.	
5	0 33 11	5
5	3 27 40	2
2	5 31 44	1
2	22 6 56	0.5
2	0 4 16	0.25
2	0 17 3	0.125
2	1 8 1	0.0625

cal significance. For example, the error for a 10-cc. sample taken at a depth of 10 cm. was calculated to be only about 0.04 to 0.05 per cent; for a 20-cc. sample taken at 3 cm. this error was about 0.08 per cent.

The concentration of the soil suspension should not exceed 2 per cent and the diameter of the sedimentation cylinder or vessel should not be too narrow. An ordinary 1000-cc. graduate cylinder is a very desirable size, although 500-cc. and 2000-cc. cylinders permit the obtaining of good results. Determinations should be made at a fairly constant temperature to minimize convection currents. This is especially true when the sedimentation time is several hours long.

It is necessary to exercise considerable caution during the removal of the sample. The pipette must not be filled so slowly that the particles in the sphere of liquid do not represent those in the theoretical thin layer at the time of sampling. On the other hand, if the pipette is filled so rapidly that there is a swirling motion of the liquid near the tip, the extracted sample will not represent the true particle distribution. A satisfactory technique of filling the

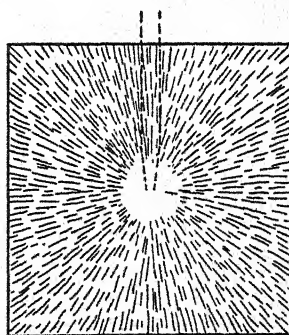


FIG. 16. Schematic drawing of the streaming of particles into the pipette during mechanical analysis (sketched after Köhn, 28). Note that the pipette removes a sphere of the suspension.

pipette is found in the use of an aspirator bottle filled with water and placed several feet above the level of the suspension. A rubber tube is attached from the top of the bottle to the pipette. Water is permitted to flow from the bottle which causes about the right amount of suction on the pipette to fill it with reasonable rapidity. In this way, all samples can be removed under the same amount of suction. The pipette must be accurately calibrated, since the amount of material in each sample must be multiplied by a rather large factor to express the total weight of particles present in the entire suspension. Small errors in an individual sample will be seriously enlarged when multiplied.

Another source of error in the weighings from each pipetteful of material extracted is the amount of dispersing agent that is present. Since sodium oxalate is so frequently used as the peptizing agent, every cubic centimeter of suspension contains a significant amount of this salt which contributes to the weight of the dried sample. The correction factor for the dispersing agent becomes more important the lower the concentration of a given fraction in the whole suspension.

Hydrometer. In 1927, Bouyoucos (5) introduced the hydrometer method for mechanical analysis. A hydrometer, which was calibrated to read grams of material per liter, is placed in the soil suspension in a special sedimentation cylinder at the desired time and readings recorded. The hydrometer was originally calibrated in actual soil suspensions of an average loam soil. From these readings an accumulation curve may be drawn. A special steamlined hydrometer for soils has been perfected by Bouyoucos (7) for sensitive readings. Theoretically, the hydrometer measures the density of a suspension at a given depth with time. Bouyoucos took readings at the end of 40 seconds to give the amount of sand (> 0.05 mm.) that had settled out. One-hour readings gave 5μ material; 2μ clay was determined at the end of 2 hours. In the earlier stages in the development of the hydrometer technique, the reading at the end of 15 minutes was taken as the "total colloidal content of the soil" since the quantity of material thus recorded was correlated with the heat of wetting, and was assumed to characterize the activity of the soil material. Bouyoucos claimed that the hydrometer method included the finer silt in its clay because of the surface activity of this silt fraction. Therefore, the technique was considered a means of distinguishing the colloidal from the non-colloidal material.

Bouyoucos (6), using Stokes' law, later suggested that the hydrometer method could be used to obtain a complete accumulation curve. Calculations showed that at the end of one hour, particles larger than 0.01 mm. in diameter had fallen $32\frac{1}{2}$ cm., the distance from the top

of the suspension "to the point where the solid soil column" was being formed. At the end of 2 hours, particles larger than 0.0071 mm. were calculated to have settled beyond the reach of the hydrometer bulb. Inasmuch as the hydrometer had to be calibrated with soils to give correct readings and since the calculations according to Stokes' law do not conform to the particle-size separations of $5\ \mu$ and $2\ \mu$ at the one- and two-hour periods, respectively, it appears that the hydrometer readings at a given time determine an average density of the whole suspension which is empirically related to definite size groupings. Bouyoucos has shown that there is a difference in concentration between the top and bottom of the suspension but that the hydrometer gives an average reading for the whole suspension.

Numerous investigators have criticized the hydrometer technique on theoretical grounds. Keen (25) has suggested that the determination has to be qualitative in nature because of the density differences from the top to the bottom of the suspension, especially since an arbitrary time is chosen to measure the average density of a layer of suspension which is continually changing with depth and time. Later he (26) pointed out that the hydrometer reading refers to a portion of the summation curve rather than a definite point. The fact that most summation curves do not exhibit sharp breaks makes it possible to use the mid-point of that portion of the curve represented by the hydrometer reading without introducing a really serious error. This suggestion would help to explain the close correlation observed by Bouyoucos and others between values obtained by the hydrometer and pipette methods. It would also clarify the discrepancies that have been reported for soils that do not possess a rather uniform distribution of various-sized particles.

Olmstead, Alexander and Lakin (39) have criticized the hydrometer technique on the basis that its calibration with a soil having a given size distribution of particles would cause it to err when used on soils with widely different size distributions. The displacement effects of the hydrometer bulb were also emphasized. They were able to obtain excellent agreement between the pipette and hydrometer methods on some soils. On other soils, the hydrometer showed much higher clay contents. They attempted to improve the hydrometer method by using the principle of the specific-gravity balance. A small plummet with a volume of 10 cc. was lowered into the suspension and weighed. It was found to check with the pipette method more often than the hydrometer; however, erroneous results were obtained with some soils.

These criticisms have been refuted by Bouyoucos, primarily on the

basis of the correlation of the results that he has obtained between the hydrometer and the pipette techniques. He has improved the sensitivity of the hydrometer which has obviated the settling of particles on the shoulder of the bulb. There is no doubt that its simplicity and rapidity make the hydrometer a valuable tool for routine mechanical analyses where extreme detail is not desired. It is now being used in numerous laboratories and thousands of determinations are being made in connection with certain engineering aspects of soils such as highway, reservoir and dam constructions.

Summary on the Fractionation of the Sample. Although it is realized that numerous satisfactory techniques may be used to determine the mechanical composition of soils, the following summary will undoubtedly prove helpful to many soil investigators:

A. DISPERSION

1. *Soils containing much organic matter.* Oxidize with 6 per cent H_2O_2 . Treat with 0.2 N HCl. Wash to remove carbonates. Peptize with 0.1 N NaOH by shaking 8 to 10 hours, or stirring 15 to 20 minutes in a high-speed stirrer.

2. *Soils containing little organic matter.* (a) Same as (1) without H_2O_2 treatment, or (b) add a saturated solution of sodium oxalate (10 cc. per 10 gm. of soil) and shake or stir as under (1).

3. *Soils high in sesquioxides.* Peptize in 0.1 N NaOH and shake or stir as under (1).

4. *Soils high in carbonates.* Remove more readily soluble carbonates by CO_2 or dilute HCl treatment and peptize with sodium oxalate.

B. FRACTIONATION

1. For routine, approximate values, the hydrometer method may be used.

2. For detailed analyses the pipette method as adapted to the centrifuge is very satisfactory.

After dispersion the sample is washed through a 300-mesh sieve. The sands (> 0.05 mm.) remaining on the sieve are dried at $105^\circ C$. and weighed.

The suspension of silt and clay is caught in a weighed bottle of about 200-cc. capacity and made up so that the suspension contains 200 gm. of water in addition to the weight of silt and clay. It is stirred for 2 minutes with a stirring machine and allowed to stand until the time for the first sampling (Table 7).

A two-way stopcock, sealed to the upper stem of a 5-cc. pipette, which is then calibrated for total volume, makes a convenient sampler.

Suction is provided by allowing water to flow out of an aspirator bottle. A cathetometer, fitted with a clamp, facilitates an accurate lowering of the pipette into the suspension. The pipette is lowered slowly with the stopcock closed until the top just touches the surface of the liquid. At the appropriate time the pipette is quickly, but carefully, lowered to the desired depth, the stopcock is opened and the pipette is filled by steady suction. As soon as the suspension reaches the base of the stopcock, the stopcock is closed and the pipette is removed. The sample is placed in a dry, weighed porcelain crucible, evaporated to dryness, dried to constant weight at 105° C. and weighed. The weight of the residue is multiplied by the appropriate factor (volume of water in suspension/volume of pipette) and expressed in percentage of the total oven-dried sample.

Other samples may be taken at the times and depths given in Table 7.

After the gravity samples are taken, the suspension is again stirred for about 2 minutes. (The suspension is practically homogeneous with respect to all the smaller particles even with the removal of several samples.)

Four 25-cc. aliquots are transferred to 100-cc. centrifuge tubes and centrifuged as soon as possible for the required time. Time is counted from the moment the motor is turned on until it is turned off, with the assumption that the times of acceleration and retardation balance. The centrifuge should be brought to the desired speed as fast as possible. A vibration tachometer is a valuable means of determining the speed. Sampling is performed as under gravity sedimentation.

If sodium oxalate is used as the dispersing agent, a correction must be subtracted from the weight of the residue from each sampling, which corresponds to the excess sodium oxalate present. If temperatures or density of particles differ from the values assumed in Table 7, equations 4 and 9 may be used to calculate the correct settling times under different experimental conditions.

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CHAPTER III

PHYSICAL CHARACTERISTICS OF SOIL COLLOIDS

In the previous chapter it was shown that the soil is a disperse system which is made up of particles of varying sizes. Evidence was presented to point out that the finer fractions were responsible for the major chemical and physical activities of the soil, because of the large amount of surface exposed per unit mass. In addition, methods for determining the mechanical composition of soils were given in order to take the soil apart and study the various mechanical separates. After having obtained a picture of the makeup of the soil and means for measuring the percentage distribution of the different particles present, it is necessary to know something about the physical behavior of the various separates before a clear understanding of the physical properties of the entire soil can be had. Since the sand and silt fractions possess such a low surface activity and serve more or less as the soil skeleton, it is essential to study the physical characteristics of the soil colloids rather thoroughly.

The colloidal fraction of soils is primarily inorganic, although varying amounts of organic colloids may be present depending upon the climate, vegetation and method of cropping. The major portion of this chapter will deal, therefore, with the physical characteristics of the inorganic soil colloids. A brief discussion of the chemical and mineralogical nature of clay was presented in the previous chapter to serve as a basis of correlation between certain physical properties and chemical constitution. It was shown that colloidal clays were primarily secondary aluminosilicate minerals with sheet-like structures or crystal lattices.

THE SHAPE OF CLAY PARTICLES

Relation between Shape and Surface. The shape of the clay particles is an important property from the physical point of view. The amount of surface per unit mass, or volume, varies with the shape of the particles. Moreover, the amount of contact per unit surface also changes with shape. It is well known that a sphere has the smallest surface area per unit volume. If such a sphere is deformed into a rod and then into a disc or plate the surface increases, with the disc or plate

exhibiting the greatest surface area. Ostwald (38) has emphasized the importance of shape in determining the specific surface of colloidal systems. A summary of several of his calculations is given in Table 8. These data are self-explanatory and point out the necessity of taking the shape of particles into consideration in all surface calculations.

TABLE 8

THE EFFECT OF SHAPE UPON THE SPECIFIC SURFACE OF COLLOIDAL PARTICLES
(AFTER OSTWALD)

Shape	Dimensions	Specific surface per cc.
Sphere	Radius = 0.6204 cm.	4.836 sq. cm.
Cube	Length of side = 1.0 cm.	6.0 sq. cm.
Disc	Radius = $\sqrt{\frac{V}{\pi h}}$	Specific surface = $\frac{2}{h} + 2\sqrt{\pi h}$
$h = 0.1$ cm. (1 mm.)	1.728 cm.	21.121 sq. cm.
$h = 0.0001$ cm. (1μ)	56.42 cm.	2×10^4 sq. cm.
$h = 0.00005$ cm. (500 m μ)	79.8 cm.	4×10^4 sq. cm.
$h = 0.00001$ cm. (100 m μ)	178.4 cm.	2×10^5 sq. cm.

In addition to the increased surface per unit volume, it should be obvious that disc-shaped particles can be arranged in more intimate contact than spherical ones. Four spheres will make six point contacts when placed in close-packed arrangement. On the other hand, four discs of the same size will make intimate contact with about three-fourths of the total flat surface area, when closely packed together. Spherical particles, therefore, do not provide favorable conditions of surface contact for great cohesion. Disc-shaped particles, however, cause high cohesion when orientation is at a maximum. In addition, laminar particles can slide over each other under an applied force, whereas spheres will permit a deformation only by a breaking down or rolling apart of the spherical arrangement. Thus, shape has a pronounced effect upon the mechanical properties of the system.

Evidence of the Plate-shaped * Nature of Clay Particles. Although many attempts have been made to develop concepts of the behavior of soils by assuming spherical particles, most of the experimental evidence indicates that the smaller fractions are distinctly non-spherical.

* Realizing that the particles are not necessarily circular, we use disc-shaped and plate-shaped interchangeably in this discussion.

This evidence is found in ultramicroscopic observations of clay sols, the double refraction of clay particles, the layering of particles during deposition and the nature of the clay crystals.

When a beam of light is passed through a colloidal system, part is transmitted and part diffracted. The diffracted light is visible to the observer and the beam can be seen in the suspension; this is the so-called Tyndall effect. Since colloidal particles are smaller than the wave lengths of ordinary light, the particles themselves cannot be seen. However, the diffracted light indicates the presence of a particle. The ultramicroscope makes possible the observing of the particles in a colloidal suspension by looking at right angles to a small portion of the Tyndall beam. When one looks at such a beam through a microscope, the particles are observed in constant Brownian movement as indicated by the paths of the spots of diffracted light.

Under the ultramicroscope, a clay sol exhibits scintillating or twinkling types of particles in that the diffracted light comes and goes and is not distinctly visible at all times. This effect is associated with non-spherical particles. Spheres and cubes will always have a long axis in the plane of observation, perpendicular to direction of incident light. Consequently, they will be visible at all times as long as they are present in that portion of the beam that is being observed. In the case of rod- or disc-shaped particles, it is necessary for the long axis to be in the plane of observation at right angles to the direction of the beam of light. Therefore, only those particles with this orientation are visible. The scintillating or twinkling effects are produced by the sudden orientation of the particles in a visible position and their equally sudden disappearance from view as they assume a different position. It is not possible to distinguish between rods and discs through this observation, although Freundlich (11) has suggested a special ultramicroscopic technique to achieve such a distinction.

If colloidal suspensions containing non-spherical particles are examined under ordinary reflected light while at rest, the observer will note a uniformity of color. If such a suspension is stirred or shaken in a rotating manner, alternate dark and light streaks, parallel to the direction of flow within the flask, will appear. These dark streaks will disappear again as the suspension comes to rest. Such an effect is caused by the orientation of the particles with their long axes in the direction of flow to give a difference in the reflection of the light. It cannot be produced with spherical or cube-like particles in which all the axes are of equal length.

Perhaps the best evidence of the disc-shaped character of clay particles is associated with the double refraction that is exhibited by

clay suspensions. Double refraction is a property of those crystals in which the velocity of light passing through the crystal is dependent upon direction; that is, the crystal allows light to vibrate with different degrees of ease in the different directions. Such a crystal has more than one refractive index. It so happens that the double refraction is closely related to particle shapes. Spherical and cube-like particles do not show double refraction since the velocity of a beam of polarized light through such crystals is independent of direction; it vibrates equally fast along all axes. On the other hand, rod- and disc-shaped particles have unequal axes, both physically and crystallographically. They have at least two refractive indices and, therefore, are double refractive.

If a suspension of colloidal clay at rest is viewed through a petrographic microscope with crossed Nicols, the field will be dark. When the suspension is rotated, however, the field becomes bright. In other words, as the particles are oriented with their long axes at right angles to the direction of light propagation, they exhibit double refraction. This phenomenon is known as **streaming double refraction**. Particles may be oriented electrically also to give **electrical double refraction**. Rotation has no effect upon singly refracting particles, such as cubes or spheres. The reader is referred to the work of Fry (12) for a complete discussion of the principle of double refraction.

It is often observed that thin layers of deposited clay curl up into minute sheets when dried. These sheets are distinctly plate-like in arrangement and are very coherent. Such a phenomenon could take place only with disc-shaped (perhaps rods) particles that settle out with their broad surfaces parallel to each other. The effect is similar to the falling of broad leaves from a tree. If a clay suspension is allowed to dry on a microscope slide and is then examined under a petrographic microscope, it will exhibit double refraction, which is further evidence of orientation during evaporation and drying.

The nature of the crystal lattice of clays suggests a disc-shaped type of particle. It has been shown that the clay minerals are constituted of alternating layers of silica and alumina sheets. This sheet-like structure of the crystal, which is similar to that of mica, would be expected to favor a plate-like shape of the entire particle, rather than a cubical, spherical or rod-like form.

From these observations, the shape of clay particles is usually considered as plate-like. The assumption of such a shape helps one to visualize the causes of the plasticity, crusting and compacting of soils. The relation of shape to these properties will be discussed in subsequent chapters.

SURFACE BEHAVIOR OF CLAY PARTICLES

Adsorption of Cations. Since many of the physical properties of soils are affected by the nature of the adsorbed ion, it is necessary to summarize the essential features of ionic adsorption and exchange. Detailed discussions of the mechanism of ionic exchange may be found in the papers of Jenny (16, 18). From a physical point of view, one is interested in the amount of ions adsorbed per unit mass (exchange capacity) and the energy with which the different ions are held on the surface (ease of replacement).

The exchange capacity of colloidal clays is dependent upon the chemical and mineralogical composition of the colloid. This fact is illustrated in Table 9. These data indicate that the montmorillonitic

TABLE 9

PHYSICAL AND CHEMICAL PROPERTIES OF SOIL COLLOIDAL MATERIAL

Colloidal property	Type of H-colloid				
	Bentonite	Lufkin clay	Putnam clay	Susquehanna clay	Cecil clay
Type of mineral.....	Montmorillonite	Montmorillonite Beidellite	Beidellite	Beidellite Halloysite	Halloysite
SiO ₂ -R ₂ O ₃ ratio.....	5.0	3.8	3.2	2.3	1.3
Exchange capacity, m.e./100 gm.	95.0	82.0	65.0	47.0	13.0
Hygroscopicity, per cent by wt. (30 per cent H ₂ SO ₄).....	21.5	20.1	18.1	15.5	6.1
Heat of wetting, cal./gm.....	15.0	13.8	11.7	5.9
Swelling, cc./gm.....	2.2	1.18	0.81	0.57	0.05
Swelling (cc.).....	2.44	1.44	1.24	1.21	0.41
Exchange capacity (m.e.)					
Hygroscopicity (per cent).....	0.22	0.24	0.28	0.33	0.47
Exchange capacity (m.e.)					
Swelling (cc.).....	10.23	5.87	4.47	4.87	0.91
Hygroscopicity (cc.)					

type of mineral has the highest exchange capacity, and the kaolinitic (halloysite) group the lowest. These relationships have been previously mentioned in Chapter II. Moreover, within the same mineralogical group there is a decrease in the total exchange capacity (milliequivalents adsorbed cations per 100 gm.) as the SiO₂-sesquioxide ratio decreases. In other words, as the content of iron and aluminum in the clay colloids increases, the total exchange capacity becomes less.

The discussion of the adsorption of cations in Chapter II (see

Table 4) showed that the valence and hydration of the ion are the dominant factors in determining the energy of adsorption and release. Divalent ions are adsorbed more strongly than monovalent. Weakly hydrated ions are held more tightly than those containing a large water hull. Jenny (16) has also pointed out that the nature of the clay minerals affects the energy with which a given ion is held. For example, the percentage release of adsorbed potassium by HCl (added at symmetry concentrations, that is, the same amount of H ions was added as there were K ions adsorbed) from permutite, Putnam clay and bentonite was 85.45, 69.55 and 50.36, respectively. The release of adsorbed hydrogen ions against KCl was 8.0 per cent, 15.0 per cent and 18.4 per cent, respectively, for the same colloidal aluminosilicates. Thus, it is seen that the nature of the clay mineral as well as the valence and hydration of the ion determine the energy with which a given ion is adsorbed on a colloidal-clay surface. The importance of ionic adsorption on the hydration and other surface behaviors of clay will be brought out later.

Adsorption of Water Molecules. THE CONCEPT OF HYGROSCOPICITY. When a sample of dry soil is placed in an atmosphere of water vapor, it will adsorb water molecules until an equilibrium is established between the soil and the atmosphere. The amount of water adsorbed increases with the clay content (specific surface), the vapor pressure of the water molecules in the atmosphere surrounding the soil and decreasing temperature. Mitscherlich (28), reasoning according to the theory of Rodewald (33), assumed that the amount of water vapor adsorbed by a soil was proportional to the total surface and attempted to calculate the surface on the basis that the water was present as a monomolecular layer. Moreover, he also attempted to distinguish between the amounts of inner and outer surface by the difference between the adsorption of water and of toluene. The toluene molecules were visualized as being too large to enter the small pores and, consequently, were only adsorbed on the outer surface. Since water was adsorbed over the total surface, the difference would equal the amount of inner surface. As a result of recent scientific contributions on the nature of clays and the hydration effects of adsorbed ions, these concepts are obviously untenable. Nevertheless, the work of Mitscherlich and Rodewald has been an outstanding step in our knowledge of the surface behavior of soils.

Rodewald developed the theory of water adsorption on the thermodynamic basis of the Clausius equation and showed that the water content of hygroscopic bodies is related to temperature and vapor pressure. The heat absorbed by a hygroscopic body at constant tem-

perature was considered as equivalent to the work used to separate the water molecules from the surface and to change liquid molecules to the vapor phase. Consequently, if heat must be applied to a hygroscopic soil to loosen water molecules and to change them into water vapor, heat must be evolved during the adsorption of water molecules if a dried soil is placed in water. This evolution of heat is known as the **heat of wetting**. Rodewald visualized the heat of wetting as being due to the total energy released by the adhesion of water to the surface. This value was then considered to be proportional to the force with which water molecules were attracted to the surface. Mitscherlich then defined **hygroscopicity** as the moisture content of the soil at which it no longer exhibited heat of wetting. To achieve this moisture content, he placed the soil in vacuo over a 10 per cent (by wt.) H_2SO_4 solution. At 25°C . this gave a relative humidity of 94.3 per cent. Values obtained by this technique were used to calculate the surface of soils.

Robinson (32) determined the amount of water adsorbed by a large number of soil colloids over 3.3 per cent H_2SO_4 (by wt.), which gave a relative humidity of 99.8 per cent. The average value thus obtained was used as a rough measure of the total colloid content, since nearly all colloids adsorbed the same amount of water at this humidity. Later work by Anderson and Mattson (3) showed that, even though colloids may adsorb almost equal amounts of water at high vapor pressures, the quantities adsorbed at lower vapor pressures vary considerably, dependent upon the nature of the colloid. They used 30 per cent H_2SO_4 (by wt.) to measure differences in the adsorptive properties of various colloids; this solution gave a relative humidity of 74.9 per cent. These investigations, similar to those of Mitscherlich, aimed to obtain an equilibrium point that could be used to express the water relations of a soil to indicate its colloidal nature. Even though a different vapor pressure was used in each of these three methods, no attempt was made to investigate the relationship between moisture content and vapor pressure.

VAPOR PRESSURE AND MOISTURE CONTENT. Thomas (35) and Puri, Crowther and Keen (31) were the first to study the relation between vapor pressure and moisture content. Determinations were made by both the dynamic and static methods. The dynamic, or air-stream, method consists of passing air of a desired humidity over or through the soil. In the static method, the soil is placed in atmospheres of known humidities until equilibrium is established. The vapor pressure is regulated by appropriate sulfuric acid-water mixtures. All vapor-pressure measurements require a rather long period of time to

establish equilibrium. At relative humidities close to saturation, the soil will continue to take up water until it is saturated. It is convenient to allow the soil to remain in contact with an atmosphere of known humidity for at least a week before measurements are made. Another factor to be considered in vapor-pressure measurements is the hysteresis effect. A soil that has reached equilibrium with a given atmosphere by drying down from a higher moisture content will always contain more water than one that has reached equilibrium by wetting or adsorbing moisture. Most experimental data, however, are obtained from the wetting curve without considering the hysteresis effect.

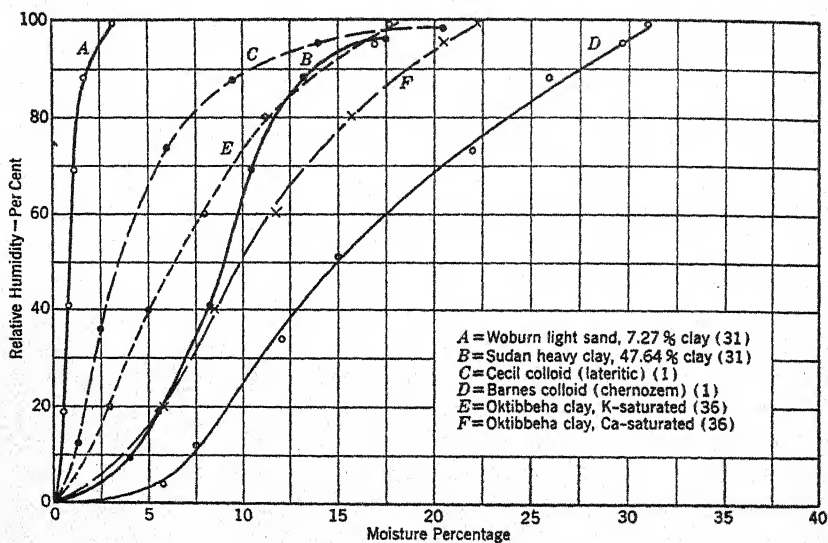


FIG. 17. Vapor-pressure curves of various soils.

These investigators found that the curves relating vapor pressure to moisture content are S-shaped, with an inflection point at about 50 per cent relative humidity. Kuron (24), as well as Alexander and Haring (1), have obtained similar curves with colloidal clays. Thomas has suggested that the relatively flat portion of the curve at low moisture contents is due to the fact that the water molecules are held so tightly that their vapor pressure is very low. The few succeeding layers of water cause large increases in the vapor pressure until finally, as the capillary spaces begin to fill, it requires large amounts of water to produce a change in vapor pressure. Typical examples of these curves are illustrated in Figure 17.

EFFECT OF AMOUNT AND NATURE OF COLLOID ON VAPOR-PRESSURE CURVES. These results in Figure 17 point out that the curves of a soil relating vapor pressure to moisture content are affected by the amount and nature of the colloid and the type of exchangeable cation on the surface of the clay. The effect of the amount of clay is evidenced in comparing curve *A*, a light sand from Woburn, with *B*, a heavy Sudan clay. At the same vapor pressure or relative humidity more water is adsorbed the higher the clay content; this is because of the greater specific surface of the clay.

The influence of the nature of the colloid is illustrated by curves *C* and *D* for the Cecil and Barnes colloids. Lateritic colloids, which contain the halloysite type of clay minerals, possess a much lower activity per unit surface than the montmorillonitic types. The Barnes colloidal material contains not only the montmorillonitic type of clay mineral but includes also organic colloids. Consequently, such material possesses a high adsorptive capacity for water molecules. Similar results for other colloids are indicated in Table 9. The activity of the surface decreases with an increase in the iron and aluminum content of the clay; this is, hygroscopicity diminishes with a decreasing SiO_2 -sesquioxide ratio.

EFFECT OF ADSORBED IONS ON HYGROSCOPICITY. The effect of exchangeable cations on the amount of water adsorbed at various vapor pressures is shown in curves *E* and *F*. The K-saturated clay adsorbs less water than the corresponding Ca-saturated system. Thomas (36) has studied the effect of various replaceable bases on the vapor pressure curves and has found that the K-, Na- and NH_4 -saturated soils adsorb less water than the H- and Ca-saturated samples at low vapor pressures. At high vapor pressures the Na-saturated soils adsorb the largest amount of water. Thomas has attributed the peculiar behavior of the sodium curve to the greater dispersion and swelling at higher moisture contents plus the possibility of the existence of hydrates of the colloidal minerals present.

The data in Table 10 point out the same general tendencies. At a relative humidity of 99.8 per cent, hygroscopicity increases according to the series: $\text{Li} > \text{Na} > \text{H} > \text{Ba} > \text{Ca} > \text{K}$; at a relative humidity of 74.9 per cent this order is: $\text{H} > \text{Ca} > \text{Li} > \text{Na} > \text{Ba} > \text{K}$. Kuron (24) has obtained similar data on a clay from Gäbersdorf; at low vapor pressures the water adsorption series is $\text{H} > \text{Mg} > \text{Ba} > \text{Ca} > \text{Na} > \text{K}$. At higher vapor pressures the Na-saturated clay adsorbs the most water. Kuron has attributed these results to the hydration of the exchangeable cations. The peculiar behavior of the Na ions is explained on the basis that at low vapor pressures the dehydrated Na-clay

TABLE 10

THE EFFECT OF EXCHANGEABLE CATIONS ON THE PROPERTIES OF COLLOIDAL CLAYS

Colloidal system	Colloidal property	Nature of cation					
		H ⁺	Li ⁺	Na ⁺	K ⁺	Ca ⁺⁺	Ba ⁺⁺
Putnam clay (beidellite)	Hygroscopicity 30 per cent H ₂ SO ₄ , per cent by wt.....	18.13	17.13	16.53	12.75	17.37	16.29
	3.3 per cent H ₂ SO ₄ , per cent by wt.....	44.76	56.12	49.22	31.12	40.91	41.95
	Heat of wetting, cal./gm.....	13.6	12.0	12.0	9.5	15.0	13.9
	Swelling, cc./gm.....	0.81	4.97	4.02	0.50	0.91	0.85
	Hydration, cc./gm. (calcu- lated from viscosities)....	5.20	5.25	5.25	4.07	5.20	5.20
	Dispersity, particles <100 mμ, per cent.....	37.4	60.1	62.2	56.3	4.9	24.4
Bentonite (montmorillonite)	Swelling, cc./gm.....	2.20	10.77	11.08	8.55	2.50	2.50
	Hydration, cc./gm. (calcu- lated from viscosities)....	35.0	28.4	24.1	21.3	24.2	34.0
	Dispersity, particles <100 mμ, per cent.....	34.0	37.8	35.2	35.7	31.2	31.9

contains a large number of pores that are too small for the entrance of water molecules. Ca-clay, on the other hand, is more open and is able to adsorb more water. At higher vapor pressures, the Na-clay swells and greater amounts of water are adsorbed than in the Ca-system.

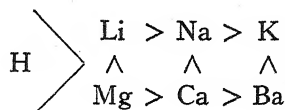
Vageler (37) emphasizes the importance of the "swarm water" associated with the hydrated exchangeable cations in the hygroscopicity of soils. The exchangeable cations are visualized as being mobile within this water layer and exerting a definite osmotic pressure. Greatest hydration of swarm ions is assumed to take place at an osmotic pressure within the water hull of about 50 atmospheres. Further hydration is explained as an association of water dipoles. Hygroscopicity is calculated on the basis of the osmotic pressures produced by the sum of all particles in the system.

It is difficult to explain, however, on the basis of pure hydration, why the divalent Mg- and Ca-clays contain more water than the Li- and Na-clays. The fact that the H-system is the most highly hydrated is even more difficult to understand. The relative hydration of the

H ion is not exactly known, although it is usually considered as not being hydrated on the basis of transference measurements.

The concept of Jenny (16), in which he considers the relative size of the ion as contributing an important part in the hydration of aluminosilicates, offers a means of understanding these data. Jenny considers, especially for permutites, that there are large cavities present "which are filled with adsorbed (wandering) ions and water molecules. For a given crystal lattice the number of water molecules which can be packed into one cavity apparently depends upon the number and size of the exchangeable ions within the cavity. The smaller the number and the true (crystal lattice radii) size of the adsorbed ions, the greater will be the number of water molecules in the system." He applies this concept to permutites with good results.

According to the Jenny hypothesis, the different clays should be hydrated according to the following scheme:



In other words, H-systems should be the most highly hydrated because of the small size of the H ion and, consequently, because of a larger cavity volume accessible to water molecules. Likewise, within the monovalent or divalent cation groups, hydration would increase with decreasing size of the ions. Since Mg and Li ions are of the same size, Mg-systems should be the more highly hydrated because there would be only one-half as many ions present in the Mg- as in the Li-systems. Similarly, Ca-systems should be more hydrated than the Na-, and Ba-systems should contain more water than the K-saturated systems. This concept has been applied to the hygroscopicity of clays and it has been found that the water content of various clay systems is inversely proportional to the volume occupied by the exchangeable cations (6).

The data in Table 9 show that hygroscopicity increases with the total exchange capacity of the colloids. The relationship is not a simple one, however, since the relative activity of the exchangeable ions and the colloidal surface must be considered in addition to the number of ions present. For example, the amount of water adsorbed per milliequivalent of exchangeable H ions increases with decreasing exchange capacity.

TEMPERATURE EFFECTS ON WATER ADSORPTION. Temperature exerts a pronounced effect upon the amount of adsorbed water at the same vapor pressure. Puri, Crowther and Keen (31) have observed

at low humidities that the relative vapor pressure of soils increases greatly with a rise in temperature from 20 to 40° C.; temperature changes within this range at high humidities have little effect upon the water content of soils. Determination of the effect of temperature on the hygroscopicity of different aluminosilicates by maintaining the vapor pressure constant at 30 mm. Hg has shown that increasing the temperature from 30 to 40° C. caused over a 50 per cent decrease in the amount of water adsorbed (7). The amount adsorbed at 80° C. was less than one-tenth as much as that at 30° C. Apparently, as the kinetic energy of the water molecules increases, less water is adsorbed by the particles.

It appears, therefore, from the numerous investigations on the adsorption of water molecules by colloidal clays that adsorption is a function of the attractive forces in the surface of the particles. These forces are associated with the chemical and mineralogical nature of the crystal lattice and the hydration of the adsorbed cations. Dehydration studies of colloidal clays have furnished some clues as to the nature of these forces.

Studies of the rate of water loss of various colloidal materials as a function of temperature have shown that water is lost at a uniformly decreasing rate up to rather high temperatures (6). For example, as illustrated in Figure 18, a permutite gradually loses water up to about 700° C. There is no break in the curve to suggest a difference in the form of water. Apparently, the last traces are held in the colloidal pores with great tenacity. Putnam clay loses water similarly as permutite up to about 350° C., when a sharp increase in water loss occurs. This increase in water loss is explained as a shattering of the crystal lattice. Bentonite continues to lose water up to about 550° C. before there is a shattering of the crystal. Most of the original water in the bentonite is lost below 150 to 200°.

Undoubtedly, adsorbed water is not completely expelled at 110° C., which is the customary temperature for removing the so-called hygroscopic water. Water that is driven off from 110° C. to ignition is generally termed combined water. This is not technically correct even though the choice of 110° C. may be analytically convenient. In light of recent results, it seems logical to consider combined water as that water which is an integral part of the crystal-lattice makeup. It is significant to note that the effect of the adsorbed cations on water adsorption and release is only apparent up to a temperature of about 300° C.

Kelley, Jenny and Brown (22) have confirmed these observations and have extended the concept to various clay minerals and soil colloids. They find that an increase in surface proportionally raises the water

content, and that this water must come from adsorption from the atmosphere in which the colloids or minerals were exposed. Most of this water is expelled below 400° C. They suggest that there are two kinds of adsorbed water. First, there is the **broken-bond water**, which is adsorbed through orientation of the water molecules (dipoles) by the free bonds of Al, Si or O ions on the exposed edges of a broken crystal lattice. This type of adsorption is considered relatively strong. Second, they define **planar water** as that adsorbed by the unsaturated valences or free electric fields on the surface of the sheets of the crystal

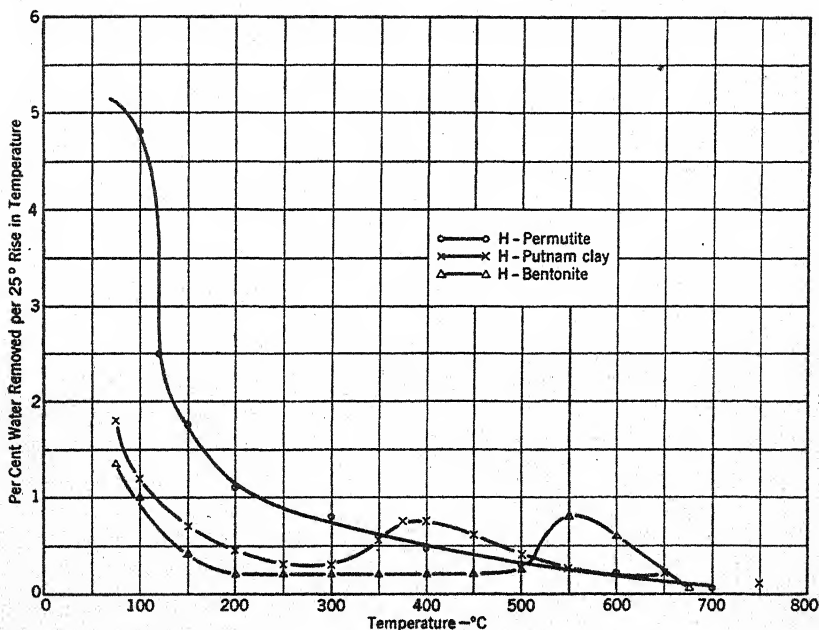


FIG. 18. Dehydration curves of permutite, clay and bentonite.

lattice. Such water is loosely held and is easily expelled at low temperatures. This type of water dominates over the broken-bond form.

These concepts offer a partial explanation, at least, of the seat of the attractive forces in soil colloids for water molecules. Even though hygroscopicity is a function of the total amount of surface, it is obvious that the amount of adsorbed water at a given humidity cannot be used as an exact measure of surface, since adsorption depends to such a great extent upon the character of the forces in the surface.

Heat of Wetting. As previously suggested in the discussion of hygroscopicity, a dry soil evolves heat when placed in contact with

water. This heat of wetting is an index of the energy of adsorption of water. Essentially, it represents the loss in kinetic energy of water molecules during adsorption. Rodewald (33) has suggested the following equation to characterize the heat of wetting particles:

$$\alpha F = E(r + i) \quad (17)$$

where i is the heat equivalent to the work necessary to diminish cohesion between particles, r the heat equivalent to the work released by adhesion of water on the surface, F the surface of the wetting substance, α the work of adhesion per unit surface, E the mechanical equivalent of heat, αF the force with which water is attracted to a unit weight of substance and $r + i$ the total energy released as heat during the wetting process.

Since the heat of wetting and heat of swelling are so closely related, the expression for the heat of swelling suggested by Katz (21) should also apply to the heat of wetting. This equation is:

$$W = \frac{Ai}{B + i} \quad (18)$$

where W is the heat produced when 1 gm. of dry swelling material takes up 1 gm. of water and A and B are constants.

FACTORS AFFECTING HEAT OF WETTING. Mitscherlich and his co-workers studied the relationship between moisture content and heat of wetting and observed a gradual lowering of the heat evolved as the soil

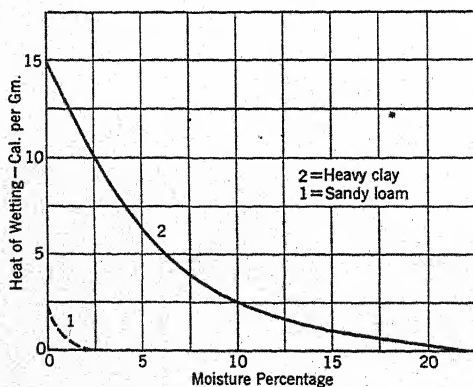


FIG. 19. The relation of the heat of wetting to moisture percentage and clay content (data of Mitscherlich, 28).

contained more original moisture. That moisture content at which no heat of wetting was obtained was then characterized as the true hygroscopicity of the substance. Their investigations pointed out also the importance of specific surface on the heat of wetting inasmuch as heavier-textured soils always evolved more heat on wetting than coarser-textured ones. These effects are illustrated in Figure 19. It is noted that the heat of wetting decreases to zero as the moisture content increases. The zero value (hygroscopicity) for the sandy loam occurs

at a much lower moisture percentage than for the clay. Moreover, curve 2 shows that clay has about 8 times the heat of wetting of the sandy loam in curve 1. Kapp (20) has shown that quartz particles larger than $5\ \mu$ do not exhibit heat of wetting. Smaller fractions show heat of wetting, but not as much as clay particles of the same size. Particles larger than $5\ \mu$ that have been fractionated from various soils evolve small amounts of heat on wetting, but only about one-tenth as much as the truly colloidal fractions.

The nature of the clay mineral and the type of exchangeable cations have similar effects on the heat of wetting as on hygroscopicity. This is necessarily so because of the interrelationship of the two phenomena. The data in Tables 9 and 10 show these effects clearly. The Li, Na and K ions cause a lower heat of wetting than H, Ca and Ba ions. Vageler (37) attributes the chief cause of the heat of wetting to the heat of hydration of the exchangeable cations. He cites the heat of hydration values of Born as follows: Li, 127 Cal. per mole; Na, 94.0; K, 75.0; Ca/2, 174.5; and H, 247. If this concept is applied to the data in Tables 9 and 10, one will find that other factors are influencing the amount of heat evolved, even though the heat of hydration of the various cations varies in the same order as the heat of wetting of the colloids.

SIGNIFICANCE OF HEAT OF WETTING. The heat of wetting has been used as a measure of the colloidal activity of the soil. The fact that the nature of the colloid as well as the amount influence the magnitude of the heat of wetting (and hygroscopicity) led Anderson (2) and others (13) to use the ratio method for estimating the colloidal content of soils. The heat of wetting, or other property, is determined for both the soil and the extracted colloid. On the assumption that the colloidal fraction contributes practically all the heat of wetting to the soil as a whole, the ratio
$$\frac{\text{heat of wetting of soil}}{\text{heat of wetting of extracted colloid}} \times 100$$
 gives the approximate percentage of colloid present.

Janert (15) has placed considerable importance upon the use of heat of wetting measurements to characterize the physical properties of soils. He has attempted to express the distribution of the soil pore space upon heat of wetting data and has presented data to show that the water content of the soil is a logarithmic function of the degree of dispersity of the pore spaces. Moreover, he has obtained a relationship between heat of wetting values and the distance that drainage tile should be spaced. In other words, he has interpreted the amount of surface per unit volume of soil, as measured by heat of wetting,

in terms of porosity. This may be possible when the amount of surface is the only factor contributing to the heat of wetting.

THE SWELLING OF COLLOIDAL CLAYS

Principles of Swelling. According to Katz (21) a solid swells when it takes up a liquid without losing its apparent homogeneity as its volume is enlarged and its cohesion is diminished. This concept, which has been developed by the botanists, attempts to distinguish sharply between capillary intake and swelling. Recently, however, greater emphasis has been placed on the forces acting in the swelling process, and the meaning of swelling has been widened to cover such amounts of pore-filling liquids as have lost their free mobility as a result of forces acting at or within the surface of the porous material. Although practically all substances that possess the capacity to swell also have the common property of a large existing or potential surface, the different physicochemical characteristics of the swelling solids, as well as of the liquid, play exceedingly important parts in the process, thereby making it extremely difficult to explain swelling by any single physical or physicochemical phenomenon. The fact that liquids used in swelling may consist of (a) associated water-like molecules with large dipole moments, (b) molecules unlike water but with dipole moments and (c) organic liquids without dipole moments suggests at least three different types of swelling. These are (a) polar liquids bound at the surface of a solid body thereby increasing its active volume, (b) liquids associated with the solid body as a solid solution (non-polar molecules) and (c) polar liquids reacting with solids to form a complex compound. The existence of these different kinds of swelling has been proved by X-ray spectography.

Swelling is always accompanied by a volume contraction of the total systems even though the dimensions of the solid material are enlarged. That is, if 1 cc. of solid is placed in 1 cc. of water, the total volume will be less than 2 cc. This volume contraction is primarily associated with the compressibility of the water during adsorption, probably as a result of oriented packing of the water molecules. There may be changes in the degree of association of water molecules. This contraction may be determined by means of a dilatometer.

Enormous pressures are developed by swelling substances. This fact was recognized by the Egyptians who used the swelling properties of wood to split rocks. Moreover, the placing of dry peas in small cavities of the trunk of a tree and then pouring water over them was an ancient custom for splitting the trunk. Posnjak (30) has studied

the relationship of the swelling pressure to the intake of liquids and has found that large pressures are developed with the adsorption of small amounts of liquid. For example, over 7 atmospheres pressure are produced when rubber takes up $1\frac{1}{2}$ times its weight of ethylene chloride. Freundlich has applied the following empirical equation to Posnjak's data:

$$P = P_1 c^k \quad (19)$$

where P is the swelling pressure, P_1 the swelling pressure when $c = 1$, c the concentration of rubber in the swollen gel and k a constant, which varies between 2.5 and 3.3 for rubber in various liquids.

Swelling of Soil Colloids. The principle of the intake of polar and non-polar liquids by colloidal material may be used as an index of swelling (41). A simple apparatus for such purposes is illustrated in Figure 20. It consists of a Jena glass tube with a fused-in, sintered-glass filter disc which is connected to a microburette by a small glass tube. A weighed amount of powdered colloid is poured into a calibrated glass tube, which is closed at the bottom with a fine silk cloth. The tube and colloid are then placed upon the moist filter disc and the intake of liquid is measured at various time intervals.

The rate of water sorption (intake) varies considerably with the nature of the colloid. It is rapid at first and then becomes slower with time. Equilibrium for ordinary colloidal clays is obtained within 1 to 3 days; for bentonites, however, sorption continues for almost a week. On the other hand, the sorption of non-polar liquids, such as benzene or carbon tetrachloride is complete within 5 minutes. The sorption of non-polar liquids increases with the dielectric constant of the liquid. This should be expected if swelling is associated with the electrical field forces on the surface of the colloidal particles. The relationship between the intake of various liquids and their dielectric constants is shown in Table 11. It is assumed that such non-polar liquids as benzene and carbon tetrachloride do not cause significant swelling and, conse-

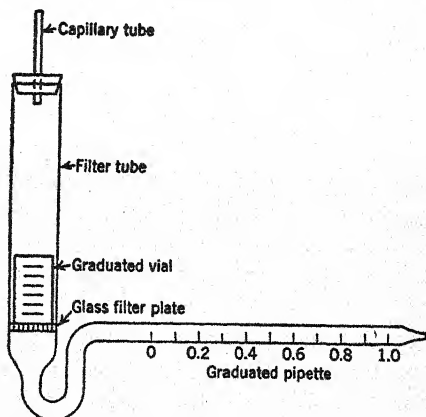


FIG. 20. Apparatus for measuring sorption of liquids by and swelling of colloidal clays.

TABLE 11

THE SORPTION OF POLAR LIQUIDS AND DIELECTRIC CONSTANTS

Liquid	Intake	Dielectric constant, D	Swelling, S	D/S
	cc./gm.		cc./gm.	
H-OH	0.99	76.0	0.58	131
H ₃ C-OH	0.66	31.9	0.25	128
H ₅ C ₂ -OH	0.60	24.1	0.19	127
H ₇ C ₃ -OH	0.57	20.5	0.16	128
H ₁₁ C ₅ -OH	0.53	14.6	0.12	122
Pore space CCl ₄	0.41	2.2

quently, may be used to measure the total pore space of the system. Part of the sorbed water fills the pores and part is oriented on the surface of the particles to produce the phenomenon of swelling. The difference between the sorption of water and the non-polar liquid, therefore, represents the amount of water that is taken up in the swelling process, although it does not represent the total volume increase of the clay because of the contraction effects previously mentioned.

It has also been shown that the swelling of colloidal clays increases with the SiO_2 -sesquioxide ratio of the colloid and varies with the nature of the adsorbed cation. The data in Tables 9 and 10 point out these effects. The expanding-lattice types of colloids (montmorillonite and beidellite) swell considerably more than the fixed-lattice types (halloysite, kaolinite). This suggests the possibility of two types of swelling, namely, intermicellar and intramicellar. Hofmann, Endell and Wilm (14) have demonstrated through X-ray techniques that the montmorillonitic clays have an expanding lattice and exhibit intramicellar swelling on wetting. The majority of the data indicate, however, that intermicellar swelling is probably the more important.

The results in Table 10 show that for Putnam clay (beidellite) swelling varies with the nature of the adsorbed cations as follows: $\text{Li} > \text{Na} > \text{Ca} > \text{Ba} > \text{H} > \text{K}$. The monovalent K ion produces the least amount of swelling. It behaves similarly to the divalent cations and the H ion. This same effect has been observed on all other clays of the beidellite type. On the other hand, the order of swelling for bentonite (montmorillonite) is: $\text{Na} \geq \text{Li} > \text{K} > \text{Ca} = \text{Ba} > \text{H}$. In the montmorillonitic colloids, the K ion occupies its normally expected place. Lutz (25) has found that lateritic colloids (halloysite)

do not swell irrespective of the nature of the adsorbed ion in the exchange complex.

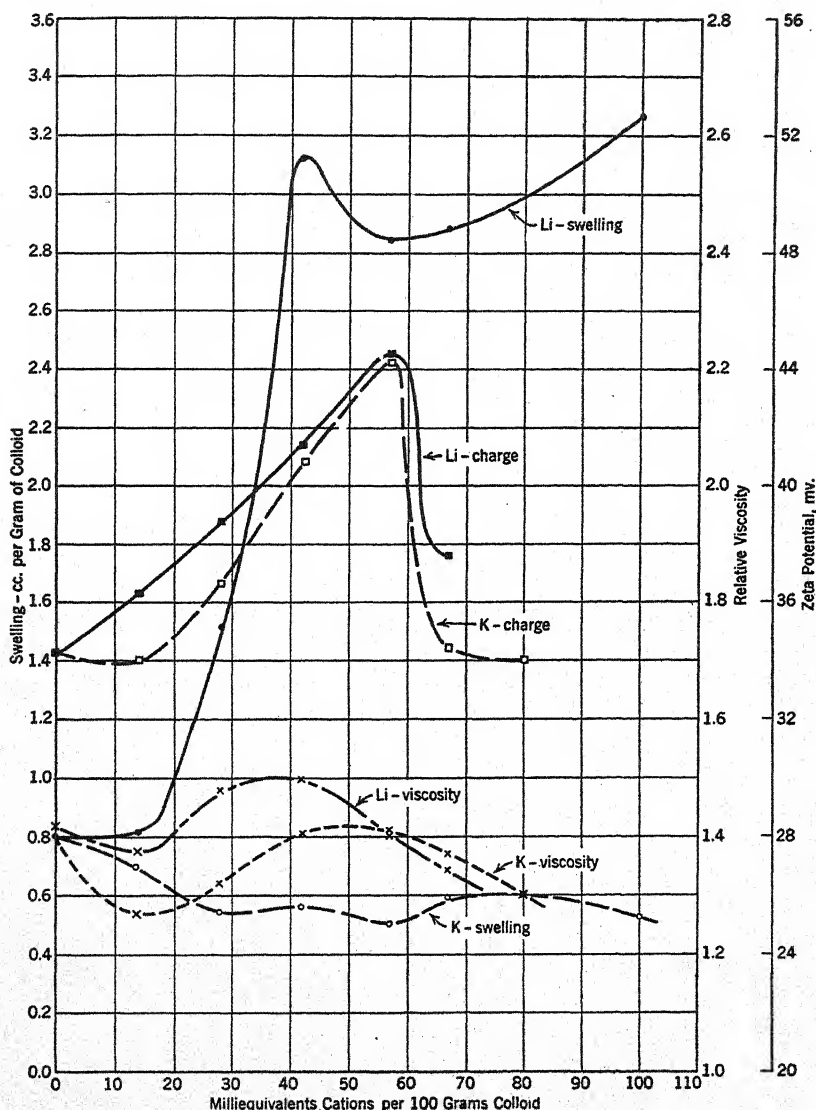


FIG. 21. Swelling, viscosity and potential of Li- and K-clays in relation to the degree of saturation with bases.

The swelling of Li- and Na-clays increases with the concentration of these ions on the complex. This is illustrated for Li-clay in Figure 21.

Maximum swelling is reached at about 60 per cent saturation of the exchange complex with Li ions. On the other hand, there is a continuous decrease in swelling as the percentage of K ions in the system increases. Differences in the strength with which these ions are held on the surface are not sufficient to explain such pronounced variations in swelling. Likewise, the fact that Li- and K-saturated clays have about the same charge and dispersity does not add to the clarity of the problem. The differences cannot be interpreted directly in terms of the hydration of the two ions. Since the K- and Li-clays behave similarly with respect to hydration in an excess of water (colloidal suspension) it appears that the K-systems are unable to orient water around the surfaces of the primary colloidal particles whenever much energy is required for the dispersion of the secondary colloidal particles. The Li-colloid attracts water molecules very strongly even to the point of complete dispersion of aggregated systems. In colloidal suspension, where the individual particles have been brought into suspension by some external force, the K-saturated surfaces do attract a rather large number of water molecules as evidenced by viscosity measurements. It seems that the dehydration of K-saturated clays causes an irreversibility of the system. The K ions apparently hold the sheet-like particles together with much more energy than Li or Na ions. This irreversibility is explained on the basis of ionic size and crystal-lattice structure.

Besides the peculiar effect of the K ion, it is interesting to note also that there is little difference in the swelling of H- and Ca-saturated colloids. In all cases, the H-system swells slightly less than the corresponding Ca-colloid. This fact will be emphasized further in the discussion on soil structure in Chapter V.

Assuming that the amount of water oriented by the exchange complex is a function of (a) the number of ions on the complex, (b) the tightness with which they are held and (c) the field strength or orienting tendency of each ion as determined by its charge and size, the relative hydration of the different Putnam clays should be as follows: K = 1.0, Li = 2.3, Na = 1.4, Ca = 1.7 and Ba = 0.9. The relative swelling for each of these systems is 1, 10, 8, 1.8 and 1.7, respectively. These data strongly suggest that there are important influences contributing to the swelling of clays other than the effect of the exchangeable ions. It seems that the forces within the inner layer can attract water molecules as well as ions to the colloidal surface. Any explanation of the effect of the different ions on hydration must take into consideration not only the orienting tendency of the ions for water molecules, the number and kind of ions present and their activities but

also the attractive power of the inner layer for water as it may be affected by ionic adsorption.

The high swelling of bentonites in comparison with colloidal clays strongly suggests that the former attract large amounts of water as a result of forces associated with the inner layer of the colloidal surface. There are several experimental facts that indicate the presence of this type of hydration. If a comparison is made of the swelling of the Li-, Na- and K-saturated bentonites and clays it is seen that these bentonite systems swell 116, 175 and 1610 per cent more, respectively, than the corresponding clays. The relative numbers of effective ions per gram of colloid as calculated from the saturation capacities and the symmetry values are 0.44, 0.47 and 0.41 milliequivalents for the Li-, Na- and K-saturated bentonites. For the corresponding Putnam clay systems these values are 0.37, 0.38 and 0.27 milliequivalents. These data suggest that such large differences in swelling in the two systems cannot be explained on the basis of ionic hydration. Moreover, the K-saturated bentonite is similar to the Li- and Na-systems in its swelling. Base exchange studies and viscosity and dispersion also point out that the different exchangeable cations, including the divalent ions, have similar effects on the dispersity of bentonite. These differences may be explained on the basis that bentonite is a montmorillonitic type of aluminosilicate with a high hydration; Putnam clay is described as a beidellitic type of colloid. Inasmuch as the colloidal behavior of bentonite classifies it as being more hydrophilic in nature than Putnam clay, it seems plausible to interpret these differences in swelling and hydration to variations in the attractive forces of the inner layer and the resulting increased mobility of the adsorbed ions, causing a measurable osmotic type of swelling. (See Mattson [27].)

✓ **Concepts of Swelling.** It is obvious that the swelling of soil colloids is a rather complicated physical phenomenon. Experimental results suggest certain interpretations of swelling in terms of the colloidal chemistry of the surface, that is, the effect of both the inner and outer parts of the electrical double layer in their relation to charge and hydration. It is interesting to compare these interpretations with the deductions of other investigators. Katz (21) has studied extensively the swelling of a large number of substances of widely varying character. Care was taken that swelling was not complicated by mechanical factors such as the porosity of the systems. From his experimental results Katz has calculated, with the aid of thermodynamical reasoning, certain relationships between swelling pressure, heat of swelling, relative vapor pressure, etc. He compares the behavior of swelling substances to that of an ideal concentrated solution (Nernst), the heat of dilution of

which can be entirely changed into other forms of energy. Evidently the swelling of soils and clays is more complicated; but there is reason to believe that the orientation of molecules on the surface of clays as a result of the electrical properties of both the liquid and the surface may follow the laws developed by Katz.

Terzaghi (34) has studied the swelling of elastic systems with porous structures (soils). He has interpreted the swelling of these systems as being due to the combined action of the surface tension of the water in the system and the elasticity of the solid components. Although the results of Terzaghi's work are of extreme importance in soil mechanics, a more physicochemical interpretation is necessary to describe the hydration phenomena in colloidal-clay systems.

Mattson (27) has explained the swelling of colloidal clays on the basis of the Donnan equilibrium as developed by Wilson and by Proctor and Wilson. Experimental results indicate that the swelling of bentonites may be at least partially explained by the Donnan concept but that this type of swelling cannot account for an appreciable amount of the hydration displayed by the common clay colloids.

The investigations on the effects of the type colloid and the nature of the exchangeable cations on swelling suggest that the concept of molecular orientation on surfaces and interfaces and around ions affords an interpretation of the process of swelling. The concept of molecular orientation has been developed by such well-known scientists as Langmuir, Harkins, Debye and others.

Volume changes in soils have a decided practical bearing. The effects of shrinkage and swelling on the structure of soils will be discussed in Chapter V.

THE VISCOSITY OF COLLOIDAL CLAYS

The Nature of Viscosity of Colloidal Suspension. The viscosity of a liquid refers to the internal friction between the molecules of the liquid. In a colloidal solution it is assumed that the particles are hydrated and that the friction takes place between the water molecules of the water hull and of the dispersion medium. Colloids are generally divided into two distinct groups, hydrophile (or lyophile) and hydrophobe (or lyophobe), with respect to their viscosity. Hydrophilic sols are characterized by a relatively high viscosity. The hydrophobic colloids do not possess a viscosity appreciably different from that of their dispersion medium. Clay sols can be considered as occupying an intermediary position, being neither truly lyophilic or lyophobic. Colloidal clay possesses the properties of hydrophilic colloids because of its

hydration. Its sensitivity to electrolytes is a hydrophobic characteristic.

Viscosity measurements have been rather extensively used for the characterization of lyophilic colloids, even though most of the work has been more or less relative in nature. It has been fairly well established that the viscosity of colloids is a function of the volume occupied by the disperse phase. This volume should include that of the particle and any water of hydration that may be associated with it. Einstein (10) has proposed a formula showing the relation between viscosity and the volume of the dispersed phase. His equation indicates that any increase in viscosity in a colloidal system depends on the total volume of the particles and is independent of the degree of dispersion. This formula is,

$$n_s = n_m (1 + 2.5 \Phi) \quad (20)$$

where n_s is the viscosity of the colloidal system, n_m the viscosity of the dispersion medium and Φ the volume of the dispersed phase per unit volume of sol.

This equation assumes that the particles are spherical and rigid. The value of Φ also includes any water of hydration associated with the particles. Meyer and Mark modified the equation to include the relationship of viscosity to the volume of the dispersed phase:

$$\frac{N_s}{N_0} = 1 + 2.5 \left(\frac{\phi}{V - \phi} \right) \quad (21)$$

where $\frac{N_s}{N_0}$ is the relative viscosity, V the total volume of sol and ϕ the effective volume of the dispersed phase. Emulsoid particles have a much larger "active volume" than that calculated from the size of the particle itself. This suggests that they are strongly hydrated. The nature of this hydration may be varied. Water may be held as a film around the particles by the forces of adsorption; it may be adsorbed within the particle; or, it may be the result of hydration of the ions of an electrolyte that are associated with the particle.

Numerous investigators have observed that the addition of small amounts of electrolytes to colloidal systems produces a sharp drop in their viscosity. Further additions either produce no change or cause an increase in viscosity, depending upon the nature of the colloid and the type of electrolyte. These viscosity changes appear to follow a course similar to that of the electrokinetic potential of colloidal systems.

Krulyt (23) calls this change in viscosity the electroviscous effect, by which is meant that the change in relative viscosity is due to the electrical charge on the particle. As this charge increases, viscosity

also becomes greater, and vice versa. Pauli (29), on the other hand, in discussing the colloid chemistry of the proteins, attributes their physicochemical behavior to their degree of ionization. Hydration of protein particles increases with ionization. Since variations in viscosity are due to differences in hydration, viscosity is a function of the degree of ionization of the protein.

Wiegner (40) considers the hydration of ions as playing the important role in the viscosity of clays. Particles containing hydrated ions on their surfaces are voluminous and viscous in pure water. Those particles containing weakly hydrated ions around the primary particle have a lower viscosity. The distribution of ions around the particle is considered as dependent upon the distribution of the different strongly hydrated ions such that those ions in the outside liquid draw water from the particle and the particle shrinks. Strongly hydrated ions within the particle draw water into the particle, producing swelling and therefore an increase in viscosity. The decrease and increase in the size of the particle are responsible for the viscosity changes.

Factors Affecting Viscosity of Clays. The effects of the amount and nature of the exchangeable cation on the viscosity of clay sols are shown in Figure 22 (4). It is evident that the monovalent and divalent cations exert entirely different effects on the viscosity of clays. These effects may be explained as representing two types of hydration changes.

Hydration of the system can be considered as being due either to a water film around the particle as produced by the adsorption of highly hydrated ions or to water held between the particles in aggregate formation. Since the viscosity of a colloidal system is a function of the "active volume" of the dispersed phase, a colloidal aggregate, containing water enmeshed between the particles constituting the aggregate, will have a larger volume than the total volume of the individual particles. In other words, the volume of the aggregate will be the sum of the volume of the particles and the volume of the occluded water. Therefore, a suspension containing aggregates will possess a higher viscosity than one containing the same number of particles in a mono-dispersed state, provided the dispersed particles have no adsorbed water hull. On the other hand, if a highly hydrated ion is adsorbed by the particle when the aggregate is dispersed, the volume of the particle will be larger, causing an increase in viscosity.

A careful study of the viscosity curves shows that both these hydration effects occur with the monovalent ions. The H-saturated clay is not a mono-dispersed system. It is composed of loosely bound, hydrated aggregates. The relative viscosity of this system is 1.4232. The addi-

tion of 14.25 milliequivalents of a monovalent base produces a marked decrease in viscosity. This decrease is greatest with the K ion and least with the Li ion. The difference between the Li, Na and K ions is undoubtedly a lyotropic effect.

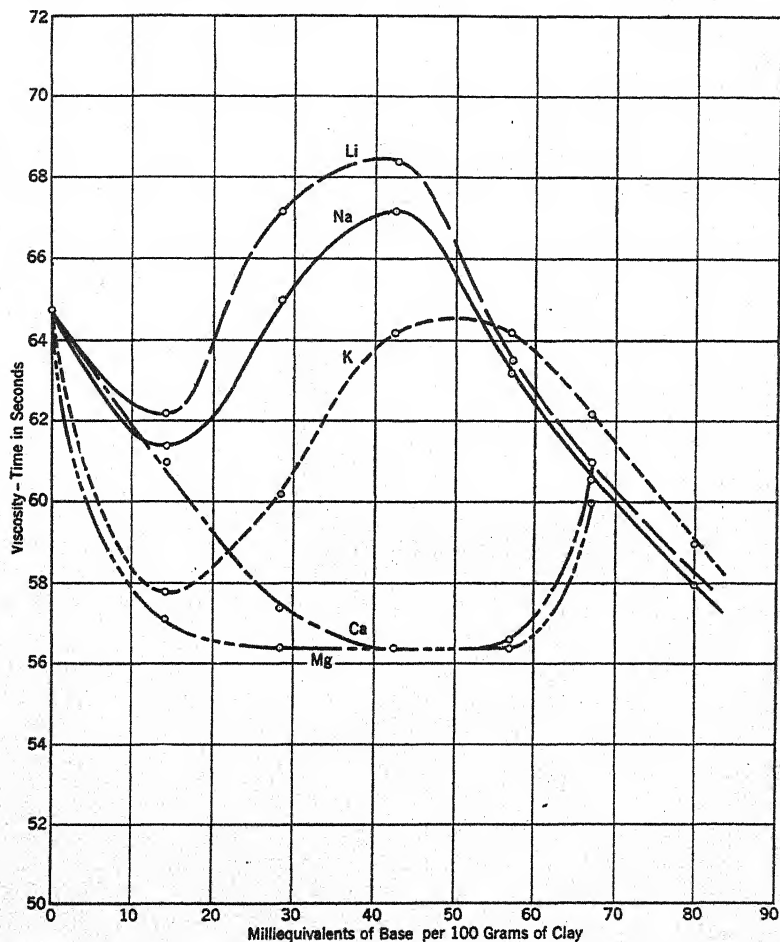


FIG. 22. The relation of viscosity of clays to the amount and nature of adsorbed cations.

There are two opposing forces exerting their influence on the viscosity changes. Dispersion of the H-clay aggregate tends towards a decrease in the total volume of the dispersed phase and a decrease in viscosity. Hydration of the dispersed particles causes an increase in this volume and in viscosity. Therefore, the K ion, with the lowest

hydration of the monovalent ions used, produces the least hydration of the particles and decreases the viscosity the greatest amount. The high hydration of the Li ion nearly counterbalances the effect of dispersion, inasmuch as the viscosity decreases only slightly. Thus, with only a slight addition of an exchangeable monovalent cation to this colloidal clay a distinct effect upon the viscosity and structure of the system is observed. The lyotropic effect of the cations is very obvious.

As the concentration of exchangeable monovalent cations becomes greater the viscosity increases to a maximum, then rapidly decreases to a minimum and increases again as flocculation begins. This initial increase represents an augmentation in hydration as more ions with a large water hull replace the hydrogen ions on the particle. Likewise, the charge on the particle and the dispersion of the system increase along with the viscosity. The decrease in viscosity from the maximum point is accompanied by a sharp diminution of the migration velocity. (See Figure 21). The maximum viscosity represents the point of largest hydration of the particle as produced by hydrated ions on the exchange complex. In other words, the exchange complex contains the largest number of cations functioning as active ions in the physico-chemical behavior of the colloid. From the standpoint of hydration, viscosity, dispersion, migration velocity and other related phenomena that are dependent upon the activity of the ions on the surface of the particle, the maximum activity should take place at this point. The colloid contains about 40 milliequivalents of exchangeable Li and Na ions and about 50 milliequivalents of K ions per 100 gm. of clay at this point of maximum viscosity. The pH value of the suspension containing these amounts of exchangeable monovalent cations is approximately 6.5.

The potentiometric and conductometric results show that the number of ions in solution, free from the surface of the colloid, increases rapidly above 50 milliequivalents of base per 100 gm. of clay. This suggests that with a larger number of ions in solution the concentration of ions on the particle exhibiting the property of hydration decreases. In other words, the Li-, Na- and K-clays begin to exhibit effects analogous to the hydrolysis of salts of weak acids. Wiegner (40) suggests that an increase in the concentration of electrolyte causes the outer layer of the colloidal particle, containing the exchangeable cations, to become tighter, decreasing the hydration of the particle. This repressing effect on the ions on the particle should be expected since conductivity measurements show an increase in the concentration of ions in solution.

The increase in concentration of ions in solution causes a corresponding decrease in hydration and viscosity to a minimum. Here an

increase in viscosity occurs, as a point is reached where slow coagulation begins. As aggregation proceeds, water is enmeshed between the particles, and this hydration of the floccule causes an increase in viscosity. The viscosity of the flocculated Li-clay is also higher than that of the Na-system, which in turn is greater than that of the K-clay. Thus, with the monovalent ions there are two types of viscosity corresponding to two types of hydration.

The divalent ions exhibit a different type of curve. There is a continuous decrease in viscosity until about 60 milliequivalents of Ca or Mg ions per 100 gm. of clay have been added. At this point there is an increase in viscosity, undoubtedly due to the slow coagulation of the sol. The Ca- and Mg-sols flocculate with the addition of 67 milliequivalents of base, which is equivalent to the saturation capacity of the colloid.

The continuous decrease in viscosity signifies a lowering of the hydration of the particles with respect to that of the original H-saturated aggregates. A diminution in the volume of the dispersed phase indicates that a replacement of the H ions on the particle by Ca ions has caused at least a partial dispersion of the original aggregate. This was confirmed by ultramicroscopic counts.

It is interesting to note that the sols containing 67 milliequivalents of Ca and Li ions have the same viscosity. However, there is a distinct difference in the type of viscosity. The Li-saturated clay owes its viscosity to water of hydration produced by Li ions on the individual particles. The Ca-sol possesses a viscosity due to water enmeshed between particles during aggregation. In other words, the distinct effect of the highly dispersive monovalent cations as compared with the divalent ions is clearly shown at this point.

Hydration and Viscosity. It is evident that viscosity changes are associated with changes in the degree of hydration of the particle. The nature of this hydration undoubtedly is a function of the ions adsorbed on the surface of the particle or of the water occluded between the particles in the case of aggregation. There is an excellent correlation between the lyotropic series and viscosity changes in most colloidal systems. Since differences in hydration are responsible for variations in viscosity, the hydration of the ions associated with the particle as well as those in the intermicellar liquid must play a significant part in the phenomena. If the hydration of ions is responsible for these effects, then the number of active ions in a system should be dependent upon the degree of ionization, or analogous phenomena, of the stabilizing ion. The ions in the non-ionized part of the system cannot be expected to exhibit hydration effects. Therefore, different colloids will show

different viscosities due to variations in the ionization of some component of the system. Various degrees of ionization and unlike hydration of ions resulting from this dissociation will determine the viscosity of the system.

The results of studies upon the changes in the viscosity and the charge of colloidal clay as produced by increasing quantities of cations indicate that the theories of Kruyt, Pauli and Wiegner are not so widely different. The viscosity and migration velocity curves confirm the results of Kruyt. A definite electroviscous effect is exhibited as the concentration of cations increases above the saturation capacity of the colloid. From Pauli's point of view, this can be explained as being due to a decrease in ionization. Or, as suggested by Wiegner, there is a decrease in the hydration of the colloid.

It is possible to use the Meyer and Mark modification of the Einstein formula (equation 21) to calculate the hydration of the suspended particles. This has been done for various colloidal clays (41, 25); the values for Putnam clay and bentonite are included with the data in Table 10. These data point out that bentonite suspensions are 5 to 7 times more highly hydrated than the corresponding Putnam clays. The bentonite surface is highly hydrated, irrespective of the nature of the adsorbed cation present. This is evidenced by the results of Marshall (26) on dispersity which show that the stability of bentonite suspensions is about the same for all ions. On the other hand, the stability of Putnam clay is dependent largely on the nature of the adsorbed ion. Therefore, the differences in the hydration of beidellitic clays and montmorillonitic bentonites, as shown by viscosity, swelling and dispersity data, are fundamentally associated with the properties of the inner layer of the crystal lattice. In addition to the influences of the inner layer there are also the hydration effects of the adsorbed ions.

SUMMARY ON THE HYDRATION OF CLAYS

Hygroscopicity, swelling, viscosity and dispersity studies permit the following generalizations concerning the hydration of clays:

1. Hydration is due to the orientation of adsorbed water molecules on the surface of the colloidal-clay particles. In the case of clays with expanding crystal lattices, this adsorption may be between the sheets as well as on the exterior of the particles.
2. Hydration is a function of the nature of the crystal lattice. Montmorillonite is more highly hydrated than beidellite; halloysite and kaolinite exhibit the least hydration. This is probably due to the fact that the montmorillonite-beidellite types of clays have

expanding lattices; the kaolinitic clay minerals have fixed lattices.

3. The chemical composition of the clay mineral also influences hydration. Within the same mineralogical class, hydration increases with the amount of silica in the crystal.

4. Hydration varies with the amount and nature of the adsorbed cations. The larger the total exchange capacity and the more highly hydrated the cation, the greater is the hydration of clay. The K ion appears to be one exception to this general statement, especially in the case of ordinary clays where it causes the lowest water adsorption and swelling. This peculiar effect is probably associated with its weak ability to orient water molecules as well as certain space relationships within the crystal lattice.

5. The hydration of clays is important in soil structure and consistency as will be discussed in subsequent chapters.

FLOCCULATION OF INORGANIC SOIL COLLOIDS

In the discussion in Chapter II on the effect of charge on the dispersion of clays, it was pointed out that clay particles are negatively charged. Moreover, evidence was presented to show that the higher the negative charge the greater is the dispersion. Data were given in Table 4 and Figures 5 and 6 to demonstrate the fact that the potential or magnitude of the charge increases with ionic hydration and decreases in general with the valence of the adsorbed ion. The conclusion was reached that a highly hydrated, monovalent ion should be on the exchange complex of the clay to obtain maximum dispersion or stability of the suspension.

Clay suspensions are rather easily flocculated by the addition of a small amount of neutral salts. In general, the amount of electrolyte required to produce flocculation decreases as the valency of the added cation increases; or, as stated in terms of Schulze's rule (39), *the flocculating power of active ions increases with their valence*. Bradfield (9) was among the first to show that this valency rule is not universally applicable to the flocculation of clays since the minimum electrolyte requirement varies with the hydrogen-ion concentration, the nature of the adsorbed cation and the concentration of the sol. These observations have been confirmed by other investigators. Jenny and Reitemeier (19), in particular, have studied in detail the significance of exchange reactions on the stability of colloidal-clay systems and have helped to clarify many of the relations between flocculation, potential and ionic exchange.

It has been demonstrated that a close relation exists between the

electrokinetic potential (magnitude of the charge) of colloidal particles and the amount of electrolyte required to flocculate the suspension (23). According to Powis (see Kruyt) it is only necessary to lower the potential to a certain value, **the critical potential**, in order to produce flocculation. Jenny and Reitemeier have studied the relation of zeta potential to (1) the charge and size of the adsorbed ion and (2) the energy of adsorption and release of these ions. Their results portray so clearly the phenomenon of flocculation of clays that the essential features of their work will be presented somewhat in detail. A portion of their results is given in Figures 6, 23, 24 and Table 4. All clay systems contained 60 milliequivalents of adsorbed cations per 100 gm. of clay.

The Effect of Charge and Size of Adsorbed Ion on the Zeta Potential. It is seen in Figure 6 that for ions of equal size the potentials of the clay particles are lower as the charge on the particle is higher. Thus Mg-clay has a lower potential than Li-clay. Ba-clay has a smaller potential than NH_4 -clay. It is necessary, however, to restrict the comparison between monovalent and divalent clays to ions of equal size because the monovalent Cs-clay has a lower potential than the divalent Mg-, Ca- and Sr-clays. Jenny formulates the role of ionic size as follows: "For rare gas type of ions of equal valency the zeta potential is higher the smaller the adsorbed cation." The magnitude of the zeta potential is a function of the average distance the ions in the outer layer (adsorbed cations) are from the inner (crystal lattice of particle). As discussed in Chapter II, hydrated ions cannot approach the negative inner layer of the particle as closely as smaller non-hydrated ions. Therefore, ions with small crystal-lattice radii become hydrated, increase the thickness of the double layer and produce high potentials, as indicated in Figure 6.

The Relation between Zeta Potential and Ionic Exchange. According to Jenny, who has formulated a kinetic theory of ionic exchange, the mechanism of such exchange may be visualized as follows:

Colloidal clay particles are plate-shaped crystals which hold on their surface adsorbed cations. Owing to heat motion and Brownian movement the adsorbed ions are not at rest but oscillate, and at times are at a considerable distance from the wall. If it so happens that on account of Brownian movement a cation of an added electrolyte slips between the negative wall and the positive oscillating ion, the electrolyte cation will become adsorbed while the surface ion remains in the solution as an exchanged ion. The more loosely an ion is held the greater is the average distance of oscillation and the greater is the possibility of replacement or, vice versa, the more tenaciously an ion sticks to the surface the less readily it will be released by the cations of an electrolyte

added to the colloidal system. The average distance of oscillation corresponds directly to the average thickness of the electric double layer, and on the basis of the picture outlined one would conclude at once that clays with high zeta potentials contain easily exchangeable ions.

The validity of this hypothesis was tested by measuring the energy of release of adsorbed ions from colloidal clays and comparing the values thus obtained with migration velocity data. The release of the adsorbed polyvalent cations was accomplished by adding KCl; that of the monovalent cations was calculated from exchange reactions with

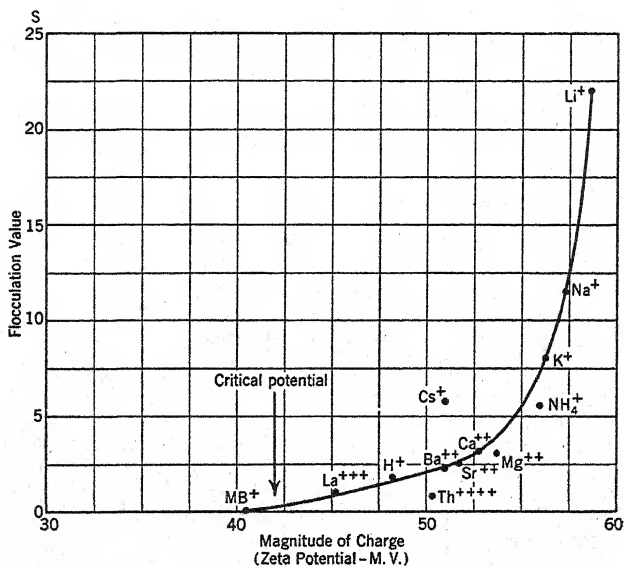


FIG. 23. The effect of the magnitude of the charge on colloidal particles to flocculation (Jenny and Reitemeier, 19). *S* denotes symmetry concentration (see text).

NH₄-clay. From the data in Table 4, it is observed that those ions that are easily exchanged cause high zeta potentials. Jenny and Reitemeier have found that ionic exchange increases approximately as the sixth power of the potentials. When the adsorbed ions are held so tightly that they cannot be readily exchanged, the zeta potential is zero.

The Relation between Flocculation Values and Zeta Potentials.

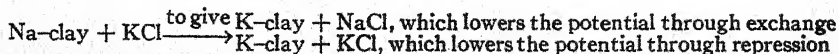
The investigations of Jenny and Reitemeier show very clearly that there is a close relationship between the stability of a colloidal system and the zeta potential. The flocculation values which were obtained by adding KCl to clays saturated with the various cations are plotted as functions of the zeta potentials in Figure 23. It is noted that the

stability of a clay suspension increases rapidly with a small rise in the zeta potential. The critical potential for the KCl-clay systems is 42 millivolts.

In light of the significance of these data to soils problems, particularly soil structure, it is essential to compare the relative positions of the H-, Ca- and Na-clays on the curve. The high state of dispersion of Na-saturated clays is easily understood by the large zeta potential. However, it is evident that the differences between acid and limed soils cannot be explained solely upon flocculation effects. This fact will be emphasized further in the discussion on soil structure.

Relation between Ionic Exchange and Flocculation. Facts have just been presented to show that flocculation of clays is intimately related to the potential of the particle and that the magnitude of the potential increases with the ease with which an ion is exchanged. It should be obvious, therefore, that flocculation is closely associated with exchange reactions. Jenny (17) recognizes two distinct situations in the flocculation of clays. First, the cation of the added electrolyte is the same as that adsorbed on the colloidal particle. In this case, flocculation results from a repression of the thickness of the double layer. The flocculation value is primarily a function of the magnitude of the original potential. This situation may be represented schematically as follows: Na-clay + NaCl $\xrightarrow{\text{to give}}$ a repression effect, which lowers the potential.

The second situation involves the addition of an electrolyte which contains an ion that is not present upon the exchange complex. In this case, flocculation results from a combination of ionic exchange and the repression effect as follows:



Jenny has found that those ions which are weakly adsorbed cause high flocculation values; those that are strongly adsorbed flocculate particles at relatively low concentrations. Using the reciprocal of the flocculation value as an index of the coagulating power of an ion, he has observed that the ability of an ion to flocculate clay increases exponentially with its adsorbability. This is illustrated in Figure 24, which also shows that the flocculating effect of a given ion depends upon the nature of the adsorbed ion originally on the clay.

These investigations point out in a concrete manner the importance of adsorbed ions on the flocculation of clays. If clays are saturated with highly hydrated cations, such as Li^+ and Na^+ , the zeta potential is high because of a thick double layer. If such a system is flocculated

with another hydrated monovalent cation, ionic exchange will be only about 50 per cent and a large amount of electrolyte (15 to 20 symmetry concentration) must be added to repress the double layer and lower the potential to achieve flocculation. This is the case if Na-clay is flocculated with LiCl. If such a system, however, is flocculated by a non-hydrated monovalent cation, such as Cs^+ , or a polyvalent ion like La^{+++} , ionic exchange will be relatively large (75 to 95 per cent) and the zeta potential will be decreased by the adsorption of a more tightly bound cation. Only small amounts of electrolyte (about 1 sym-

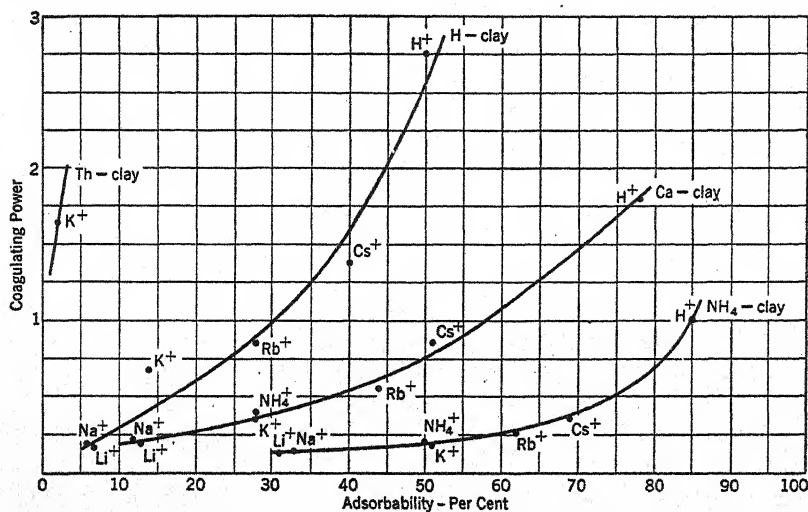


FIG. 24. The relation of the flocculation of clays to adsorption (Jenny and Reitemeier, 19).

metry concentration) are required to repress the new double layer and to produce flocculation.

If clays containing polyvalent cations, or non-hydrated monovalent ions, on the exchange complex are flocculated with electrolytes containing hydrated monovalent cations (NaCl , LiCl), ionic exchange will be only about 25 per cent. This will tend to increase the potential and stability of the system. However, the repressing effect of the added electrolyte soon offsets the peptizing influence due to ionic exchange and flocculation results. Two to ten symmetry concentrations of electrolyte are usually necessary to produce flocculation. On the other hand, if these clays are flocculated by electrolytes containing highly adsorbed cations, flocculation will take place at concentrations less than symmetry.

Finally, Jenny explains the dispersing action of the OH ion, which has puzzled many investigators, on the basis of ionic exchange. The cation of the added base causes a replacement of more strongly adsorbed ions which increases the potential. If the adsorbed ions are H^+ , its activity is nullified through the formation of water. If the adsorbed ions are Ca^{++} and Mg^{++} , the concentration of the resulting hydroxide is too small to cause flocculation.

THE PHYSICAL PROPERTIES OF ORGANIC COLLOIDS

Since the organic fraction of some soils is such a paramount factor in affecting the physical properties of the soil as a whole, several of the more important properties of organic colloids will be discussed briefly.

Viscosity and Hydration. The viscosity of colloidal humus is closely related to the nature of the adsorbed ions (5). H^- , Ca - and Ba -saturated systems have higher viscosities than the corresponding Na^- , Li - and K -humates. The former are slightly flocculated and viscosity is partly due to occluded water. There is little difference in the viscosities of the monovalent humates. By the use of the Meyer and Mark equation, Na -humate is calculated to be about $1\frac{1}{2}$ times more highly hydrated than Na -Putnam clay. It is not as hydrated as Na -bentonite. Ionic exchange reactions also suggested that humus is intermediate between clay and bentonite with respect to its hydration.

Further evidence of the hydration of the colloidal humus saturated with monovalent cations exists in the inability to dehydrate the humus sufficiently with alcohol to produce flocculation. The amount of alcohol required to remove occluded water in the slightly flocculated systems increases according to the order: $H > Ca > Ba$.

Charge and Flocculation. Humus is a negatively charged colloid. The migration velocity of various humates increases according to the ionic series: $Li = Na > K > H > Ca > Ba$. The combined effects of hydration and charge apparently make it very stable since flocculation of Li^- , Na^- , and K -systems was not achieved either by the addition of electrolytes containing monovalent cations or by alcohol. The flocculation effects of various cations decrease as follows: $La^{+++} > H^+ > Ca^{++} > Sr^{++} > Mg^{++} > K^+ > Na^+ = Li^+$. The Mg ion acts more like the monovalent cations with respect to flocculation than like the other divalent cations. This effect may be important in connection with the formation of Mg -solonetz soils.

As previously suggested, the stability of the various humates appears to be largely a function of hydration. The amount of alcohol that is required to produce flocculation by dehydration is dependent

upon the potential of the humus particle. The potentials of the monovalent humates are too high for alcohol to be effective.

Reversibility. Boutarie and Thevenet (8) have found that dried humus cannot be resuspended in water. Other studies (5) confirm this conclusion and show further that reversibility is related to the nature of the ion adsorbed on the humus. K-humus is completely reversible within a period of two days. The reversibility of Ca- and H-humates is very slow. H-humus, which is dried at 50° C., is only about 11 per cent reversible after 10 days; a 50 per cent Ca-saturated humus is 22 per cent reversible during the same time. The reversibility of Ca-humus during a 10-day period dried at 110° C. is about 4 times that of H-humus dried at the same temperature. These results point out the importance of dehydration in affecting the properties of colloidal humates.

As a general summary, it seems that the properties of organic colloids are affected by the nature of adsorbed ion in much the same manner as the inorganic colloidal material. The hydration of the ion plays an important role in increasing the hydration of the humus. Colloidal humus has an exchange capacity of about 400 milliequivalents per 100 gm. as compared with 60 to 100 milliequivalents with clays and bentonites.

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CHAPTER IV

SOIL CONSISTENCY

DEFINITION AND SIGNIFICANCE

Soil consistency is usually defined as a term "to designate the manifestations of the physical forces of cohesion and adhesion acting within the soil at various moisture contents. These manifestations include (1) the behavior toward gravity, pressure, thrust and pull, (2) the tendency of the soil mass to adhere to foreign bodies or substances, and (3) the sensations which are evidenced as feel by the fingers of the observer (24)." This definition implies that the concept of soil consistency includes such properties of the soil as resistance to compression and shear, friability, plasticity, stickiness. All these properties are manifested differently as the forces of cohesion and adhesion within the soil mass vary. To the field observer, these properties find expression through the sensation of feel. Experience has taught him to recognize the significance of these sensations in terms of differences in soil behavior.

Field observations and experimental investigations have pointed out that the consistency of soils varies with texture, organic matter, the amount and nature of the colloidal material, to a certain extent with structure and especially with the moisture content. The technical significance of these factors will be discussed in subsequent paragraphs. It is essential at this time, however, to call attention to the variations in consistency as related to the moisture content.

It is a well-known fact that soils exhibit flowage at high moisture contents; they are viscous. At lower water contents, the soil no longer flows; it is stiff, sticky and plastic. As the moisture content is decreased further, the soil loses its stickiness and plasticity and becomes crumbly or friable; it feels soft to the touch. Finally, in the dry state, it is hard and coherent; the sensation of harshness is experienced when the soil is rubbed between the fingers.

Atterberg (1), who was among the first to attach much significance to soil consistency, describes the changes in consistency with moisture as follows (26):

When one mixes clay in a powdered condition with much water, one gets a flowing clayey paste. With less water the paste becomes thick flowing. Through evaporation of the water, the clay goes over gradually into a tough, sticky mass that sticks to the fingers, wood, and metals. On account of the smeary, sticky condition the clay is not yet considered to be plastic. With further drying the stickiness disappears. The clay now can be easily molded without sticking to the fingers, and has now that which is required for "normal consistency" for the clay-working industry and is recognized as plastic and "just right for the hand." With yet further drying out the formability comes finally to an end. The clay is indeed yet moist, the clay pieces can now, however, only be united by pressure to form a mass that hangs together. Finally it loses even this formability.

The clay shows therefore at different water contents very different properties. Sometimes they are flowing or half-flowing, sometimes they are sticky and smeary. Sometimes the clay will form well and roll out well, and sometimes it is only formable through pressure. Finally, when the clay has completely dried out, it becomes firm and hard.

FORMS OF SOIL CONSISTENCY

Atterberg (2) recognized six consistency forms. They are (1) the thin-flowing consistency, when the soil flows as easily as water, (2) the thick-flowing consistency, when the soil is paste-like and flows under the force of gravity, (3) the tough-flowing consistency, when the soil is like a thick paste and does not flow under the force of gravity, (4) the plastic consistency, when the soil can be molded, (5) the loose or friable-firm (*losere feste*) consistency, when the soil granules tend to hang together as they are slightly compressed, and (6) the more hard firm consistency when the soil contains an insufficient amount of water to cause it to hang together.

These suggestions of Atterberg can be condensed to give four essential forms (not including the viscous state) of consistency that most soils may be expected to exhibit. They are (1) the sticky consistency, as evidenced by the property of stickiness or adherence to various objects, (2) the plastic consistency, as manifested by the properties of toughness and the capacity to be molded, (3) the soft consistency, as characterized by friability, and (4) the harsh consistency, which has the pronounced characteristics of hardness. The relation of these consistency forms to soil-moisture content is shown schematically in Figure 25.

At low moisture contents the soil is hard and very coherent because of a cementation effect between the dried particles. Clods will be

produced if the soil is tilled in this condition. As the moisture content increases, however, water molecules are adsorbed on the surface of the particles, which decrease the coherence and impart friability to the soil mass. The zone of friable consistency represents the range of soil moisture in which conditions for tillage are at an optimum. As the

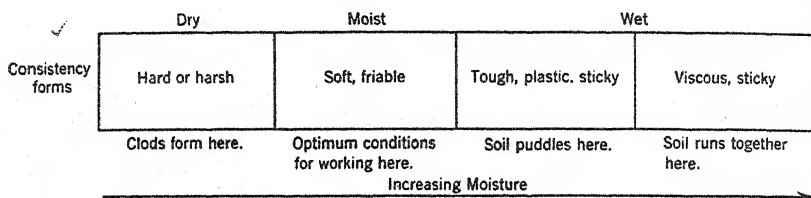


FIG. 25. Soil consistency variation with moisture.

amount of water in the soil is augmented further, the cohesion of water films around the particles causes the soil to stick together and be plastic. Soils are easily puddled in this moisture range. Some soils exhibit stickiness within the plastic range; others do not become sticky until they approach a viscous consistency.

CONSISTENCY OF DRY SOILS

A thoroughly dried soil with normal compaction usually exhibits a decided hardness or coherence in the field. The extent of this coherence naturally varies with the structure of the soil, since porosity determines the number of particles per unit volume. This in turn correlates with the amount of surface contacts and the coherence of the soil mass. Measurements of the consistency of dry soils are usually based upon the coherence of dried briquettes. The soil is wetted, kneaded and molded into rectangular briquettes (or in a cylindrical form) and dried. The tenacity or coherence of the briquette is then determined by crushing or by breaking. Crushing strength, or the "Druckfestigkeit" of Puchner (22), is measured by the force required to crush the briquette, usually in a direction parallel to its long axis. The tensile or breaking strength, or the "Festigkeitzahl" of Atterberg (2), is the force required to break the briquette through the middle. Various techniques may be employed to crush or break the briquettes.

Schübler (see Table 1) was the first to show that the breaking strength or tenacity of dried soil is an important physical property and is affected by the clay and humus content. Later investigations by Puchner and Atterberg pointed out the significance of the coherence

between clay particles as the primary cause of high resistances to crushing or breaking. Puchner gave the following values for the coherence in grams of different soil separates: 0.25–0.01 mm., 200 gm.; 0.01–0.005 mm., 35,000 gm.; 0.005–0.0015 mm., 58,000 gm.; and less than 0.0015 mm., 96,500 gm. Hardy (14) found that siliceous soils have a much greater resistance to crushing than lateritic, even though the clay and silt contents of both groups are about the same. Winterkorn (30) reported the following results for the tensile strength of three H-saturated soils: Cecil clay, with 77 per cent $5\ \mu$ particles, 61 lb. per sq. in.; Hagerstown silt loam, with 42 per cent $5\ \mu$ clay, 105 lb. per sq. in.; and Putnam silt loam with 31 per cent $5\ \mu$ clay, 165 lb. per sq. in. In spite of the high clay content of the Cecil soil, its lateritic nature is responsible for the lowest coherence.

These facts may be explained by the differences in the surface properties of colloidal clays as discussed in Chapter III. The lateritic colloids, which are composed primarily of kaolinitic clay minerals, possess a low surface activity. Since only weak attractive forces exist in the surface, a low coherence between particles should be expected. Undoubtedly, the particle shape is an important factor in the coherence of dried briquettes. Plate-shaped particles can be oriented to give a close packing during the kneading of the briquettes. These particles, with their high specific surface and greater contact per unit surface, should cohere tenaciously, as experimental observations indicate.

The coherence of dried soils, therefore, may be visualized as depending upon the amount of surface contacts per unit volume of the soil mass and the magnitude of the attractive forces within the surface. In the case of dried particles, coherence takes place in the absence of water molecules on the surface. The attraction is between solid particles. This is evidenced by the fact that the addition of small amounts of water, to form a thin layer of water molecules on the surface of the individual particles, causes a decrease in coherence and imparts friability to the soil mass.

CONSISTENCY OF MOIST AND WET SOILS

Friability. Friability characterizes the ease of crumbling of soils. That moisture range in which soils are friable is also the range in which conditions are optimum for tillage. Soils are usually in good tilth when they are friable and mellow. The individual granules are soft; cohesion is at a minimum. There is sufficient moisture between the individual particles to minimize the cementation effects that are dominant in the zone of harsh consistency. On the other hand, there

is not enough water present to cause the formation of distinct films around particle contacts to produce the cohesion that exists in the plastic range. The crumbs are probably held together, in part at least, by the orientation of water molecules between the individual particles in a manner similar to that suggested by Russell (23). The dipole nature of water makes possible such oriented adsorption.

Various methods have been proposed for obtaining a measure of friability as an index of tilth. These will be discussed in conjunction with tillage in Chapter IX.

Plasticity. It has been shown that soils (excepting those that are non-plastic, such as sands) become plastic as the moisture content increases. They are tough and exhibit considerable cohesion; they can be molded like putty. Plasticity is a property of clays that was recognized in ancient times. Modifications of its original meaning are used by modern soil investigators and ceramists. Mellor (18) defines plasticity as "the property which enables a clay to change its shape without cracking when it is subjected to a deforming stress." Wilson (29) visualizes plasticity as that property of a material which enables it to be deformed without rupture when the material is subjected to force in excess of the yield-value. ✓

These definitions imply that plasticity is a characteristic of clay to take up water and to form a mass that can be deformed into any desirable shape, and which will maintain this shape after the deformation pressure is removed. Moreover, the shape will remain unchanged after the water is removed. Sands can be molded when they are wet, but the molded form falls to pieces when it is dried; consequently, sands are not plastic.

PHYSICAL ANALYSIS OF PLASTICITY. Mellor has analyzed plasticity as being a mechanical property which is the resultant effect of a pressure and a deformation. The extent of the deformation for a given system is determined by the distance the particles can move without losing their cohesion. The pressure that is required to produce a specific deformation is an index of the magnitude of the cohesive forces that hold the particles together. These forces vary with the thickness of the water films between particles. Since the amount of deformation that can be produced varies with the size and shape of the particles, it is evident that the amount of surface present determines the number of water films contributing to cohesion. Thus, it appears that plasticity is a property which expresses the magnitude of the film forces within the soil and the effects of these forces in determining the extent to which the shape of the soil mass can be permanently changed without breaking.

LAWS OF PLASTIC FLOW. The essential difference between viscous and plastic flow is that a certain amount of stress must be added in the case of the latter before flow is produced. The volume of flow, as a function of the force applied, is characterized by the familiar Bingham equation:

$$V = k\mu(F - f) \quad (22)$$

where V is the volume of flow, μ is the coefficient of mobility, F the force applied, f the force necessary to overcome the cohesive forces of the system and just enough to start the flow (this force is the so-called "yield value") and k is a constant.

It is obvious that equation 22 may be used to characterize viscous flow when $f = 0$. Then the volume of flow is proportional to the force applied and the coefficient of viscosity of the liquid.

The difference between viscous and plastic flow is illustrated in Figure 26. Curve OE shows how viscous flow increases directly with the applied pressure. Curve $OABC$ indicates that a certain pressure must be applied to plastic bodies before flow is started. Finally, flow is proportional to the applied force, as indicated by the segment BC . The yield value is obtained by extrapolating the segment BC to the point D on the

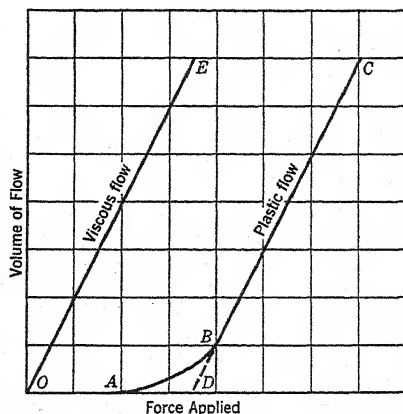


FIG. 26. A comparison of viscous and plastic flow.

abscissa. The magnitude of the yield value is correlated with the extent of the cohesive forces of the water films between particles.

Cohesion and Adhesion. It is necessary to distinguish between cohesion and adhesion in an analysis of the causes of plasticity. **Adhesion** refers to the attraction of the liquid phase on the surface of the solid phase. Thus, water molecules may adhere either to the surface of soil particles or to objects brought into contact with the soil. **Cohesion** in wet soils takes place between the molecules of the liquid phase that exist as bridges or films between adjacent particles.

WATER FILMS AND COHESION. Haines (13) has developed a theoretical concept of cohesion in an ideal soil on the basis of the surface-tension forces which arise from the water films between particles. The ideal soil is visualized as consisting of uniform spheres with radius a ,

that are arranged either in an open- or a close-packed state. In open packing or cubical arrangement (Figure 27a) there are six points of contact per particle. In the close-packed, or tetrahedral grouping (Figure 27b), there are twelve contacts per sphere.

At low moisture contents, most of the film water is found as annular rings around the various points of contact, as illustrated in Figure 27c. Each of these films tends to draw the particles together. The total

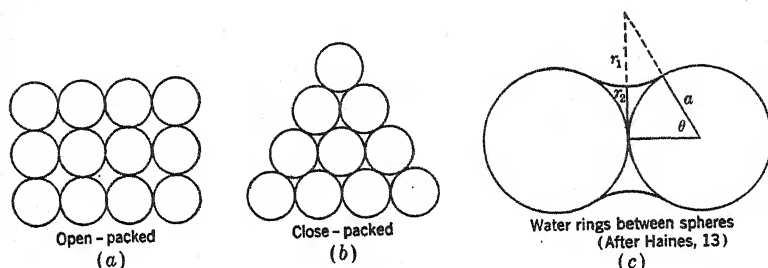


FIG. 27. Schematic representation of the packing of spheres in relation to porosity (a, b), and the nature of water rings between spheres (c).

cohesive force produced by all the films is equivalent to the summation of the individual forces exerted at each point of contact.

If r_1 and r_2 are the radii of curvature of the moisture film in each of its two principal directions and T is the surface tension of the liquid, then the pressure deficiency inside the meniscus is:

$$p = T \left(\frac{1}{r_1} - \frac{1}{r_2} \right) \quad \text{or} \quad \frac{T(r_2 - r_1)}{r_1 r_2} \quad (23)$$

where $\frac{1}{r_1} - \frac{1}{r_2}$ is equal to the sum of the two curvatures at right angles to each other. (This value is sometimes erroneously called the total curvature.) The tensile strength of the film due to the pressure deficiency inside the liquid is equal to $\pi r_2^2 p$, or $\pi r_2^2 T \left(\frac{1}{r_1} - \frac{1}{r_2} \right)$.

Fisher (12); in analyzing the theoretical considerations of Haines, has shown that the total tensile force of the water films is equal to the pressure inside the meniscus plus the tension exerted by the air-water interface; the latter is equal to the circumference of this interface times the surface tension, or $2\pi r_2 T$. The total tensile force is given by the formula:

$$F = \pi r_2^2 T \left(\frac{1}{r_1} - \frac{1}{r_2} \right) + 2\pi r_2 T = \pi r_2 T \left(\frac{r_2 + r_1}{r_1} \right) \quad (24)$$

A geometrical solution of the values of r_1 and r_2 (Figure 27c), in terms of the radius of the particle, may be obtained as follows:

$$r_1 + r_2 = a \tan \theta$$

$$r_1 = a (\sec \theta - 1)$$

$$r_2 = a(1 + \tan \theta - \sec \theta)$$

By substituting these values in equation 24 and reducing,* the following formula is obtained:

$$F = \frac{2\pi a T}{1 + \tan \frac{1}{2} \theta} \quad (25)$$

It is seen from this equation that the total tensile strength becomes $2\pi a T$ as θ approaches zero. This limit cannot be reached, however, since equation 24 breaks down as r_1 and r_2 approach molecular dimensions. Cohesion is the sum of these individual film forces over unit cross-sectional area.

When equation 25 is transformed to give the tensile stress per unit cross-sectional area within the soil mass, it is found that the total pull per unit cross section varies as T/a . Therefore, cohesion varies directly as the surface tension and inversely as the radius of the particle. According to Haines, the validity of this formula breaks down when water fills more than one-fourth of the pore space, because of the coalescence of the films. High values for maximum cohesion within a soil depend upon small particles. This means that cohesion is a function of the number of films.

Nichols (20) has applied the concept of the cohesion of moisture films to plate-shaped particles. He has based his ideas on the explanation by Duff (11) of the nature of the film action between plates.

* The following substitutions make this simplification possible:

$$\begin{aligned} F &= \pi a T \frac{(1 + \tan \theta - \sec \theta) (\tan \theta)}{\sec \theta - 1} \\ &= \pi a T \frac{\sin \theta (\cos \theta + \sin \theta - 1)}{\cos \theta (1 - \cos \theta)} \\ &= \pi a T \frac{\sin \theta (\cos \theta + \sin \theta - 1)}{\cos \theta (1 - \cos \theta)} \frac{(1 + \cos \theta) (1 + \cos \theta + \sin \theta)}{(1 + \cos \theta) (1 + \cos \theta + \sin \theta)} \\ &= \pi a T \frac{2 (1 + \cos \theta)}{1 + \cos \theta + \sin \theta} \quad \text{or} \quad \pi a T \frac{2}{1 + \frac{\sin \theta}{1 + \cos \theta}} \quad \text{or} \quad \frac{2\pi a T}{1 + \tan \frac{1}{2} \theta} \end{aligned}$$

The cohesive force of a water film between two particles is given by the formula:

$$F = \frac{k4\pi rT \cos \alpha}{d} \quad (26)$$

where k is a constant, r the radius of the particles, T the surface tension, α the angle of contact between the liquid and the particle (generally assumed to be zero) and d the distance the particles are apart.

The cohesive force should vary inversely with the moisture content for a given size and number of particles. Moreover, as long as no excess water is present, the product of the cohesive force and moisture content should be a constant. Nichols has verified this fact experimentally with synthetic soils, as shown by the data in Table 12. It

TABLE 12

RELATION OF COHESION TO THE MOISTURE CONTENT OF SOILS (NICHOLS, 20)

Soil	Moisture content (D), per cent	Cohesion (F), gm. per sq. in.	$F \times D$
Sand $\frac{2}{3}$, clay $\frac{1}{3}$	10.90	17.25	188
	12.90	15.00	193
Sand $\frac{1}{3}$, clay $\frac{2}{3}$	12.73	26.40	336
	13.10	22.50	294
Pure Cecil clay.....	13.55	56.00	759
	17.50	49.00	857

is noted that cohesion increases with the clay content, but decreases with the amount of moisture. The latter is true only because the lower moisture values in this table represent the points of maximum cohesion on the moisture content-cohesion curves. Both Haines and Nichols have shown that cohesion increases up to a maximum and then decreases rapidly as the moisture content of the soil is raised.

These facts are readily explained on the basis of equations 25 and 26. Both these formulae suggest that cohesion is a function of the number and thickness of the films. In equation 25, cohesion decreases as Θ increases; in equation 26, cohesion becomes less as d enlarges. For a given value of Θ or of d , the magnitude of the cohesion will depend upon the summation of the individual film forces, that is, the

total number of films. Consequently, maximum cohesion is obtained at that moisture content at which a moisture film is established at all points of contact. As the thickness of the moisture film increases beyond this limit, cohesion decreases. The number of films is dependent upon the colloid content. Clay soils, therefore, exhibit higher cohesive forces than sands. Plate-shaped particles produce greater cohesive effects than spherical ones.

VARIATION OF COHESION WITH MOISTURE. Atterberg and his co-workers (2, 15) were among the first to study the changes in the cohesive properties of soils as a function of moisture. The technique consisted in working the soil into briquettes, drying them to various moisture contents and determining the force necessary to cause a metal wedge to penetrate the mass. Typical results of Johannsen are illustrated in curves *A* and *B* in Figure 28. Cohesion increases with decreasing moisture content. Clay has a much greater cohesion than the fine sandy loam. There are two distinct portions to each of these curves.

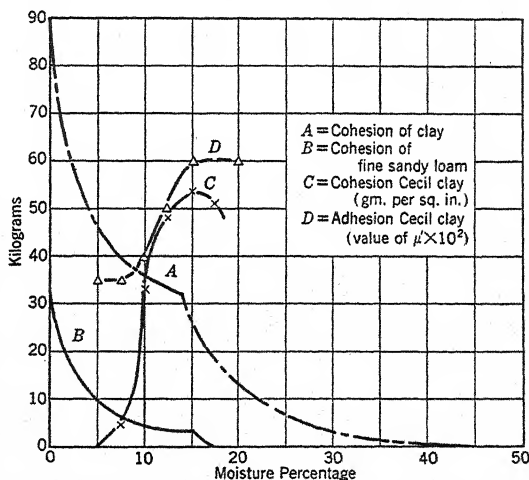


Fig. 28. The relation of cohesion and adhesion to soil moisture. Curves *A* and *B* are taken from Johannsen (15); curves *C* and *D* are from the data of Nichols (19, 20).

to light at the break point in the curve. They considered this break to represent the limit of shrinkage of the soil and the entrance of air into the pores, which were originally filled with water.

Cohesion at moisture contents above this point is due primarily to film forces; below this point the molecular attraction between the surfaces of the colloidal material is the dominant factor in cohesion. Thus, the cohesion of the fine sandy loam increases only slightly with decreasing moisture below the break point; the cohesion of clay, however, increases rapidly at low moisture contents. It should be kept in mind in the interpretation of these curves that cohesion was measured on puddled soils that were permitted to dry. This technique

produces maximum contact between particles which causes high cohesion in the dried states.

Nichols (19) has studied the relation of cohesion to the moisture content of non-puddled soils by measuring the force necessary to pull a compressed soil apart. His results are very enlightening from the point of view of the film concept of cohesion. Curve *C* in Figure 28 shows the variation in the cohesion of Cecil clay as the moisture content is increased. The loose, dry soil has no cohesion. As water films are formed, however, cohesion increases rapidly to a maximum and then decreases. The decrease represents the loss in cohesion that results from a thickening of the water films between particles. Maximum cohesion increases with the clay content of the system. The importance of these results to tillage problems will be discussed in Chapter IX.

WATER FILMS AND ADHESION. The adhesion of a foreign object to a soil should take place only at moisture contents above that of maximum cohesion. At these higher moisture contents, the water is held less tightly by the particles and is attracted on the surface of the object to form connecting films between it and the soil. The adhesion of the soil to the object is through the medium of these films. The moisture content at which maximum adhesion occurs depends upon the amount of water required to satisfy the films between the individual particles and the attractive forces in the surface of the foreign object. Nichols has observed that the force of adhesion of soil to metal is a linear function of the colloid content.

The moisture content of maximum adhesion is uniformly higher than that for maximum cohesion in the same group of soils. This is illustrated by curve *D* in Figure 28. The adhesion and cohesion curves are S-shaped; the former is located slightly higher on the moisture scale as should be expected from the film theory of cohesion and adhesion.

SOIL PLASTICITY

Plasticity should be expected to vary with particle size and shape, since it is a surface phenomenon associated with water films. Coarse particles exhibit no plasticity, but numerous data are available to show how plasticity increases with the content of smaller particles. Moreover, Atterberg (1, 2) was among the first to point out that plate-shaped particles are the most plastic. These two facts simply mean that plasticity is a function of the amount of surface and the number of contacts per available surface. Consequently, cohesive forces between particles are greater because of increased film ten-

sion; this has been explained in the discussion on cohesion and adhesion.

It follows, therefore, that the amount and nature of the colloidal material determine plasticity to a considerable extent. The colloidal material acts as a lubricant between coarse particles and diminishes their friction. In addition, colloidal material increases film tension as a result of the greater surface exposed. Any explanation of soil plasticity, therefore, must include an evaluation of the surface properties of soil colloids.

Film Theory of Soil Plasticity. If the colloidal material in soils acts as a lubricant between the coarser particles, it is highly probable that the plate-shaped colloids are oriented in such a way that their flat surfaces are in contact. This orientation increases the amount of contact between the colloidal particles. The increased contact, together with an increase in the proportion of water-film surface to the particle mass, may be considered as producing the plastic effects. In other words, within a certain moisture range, the tension effects of

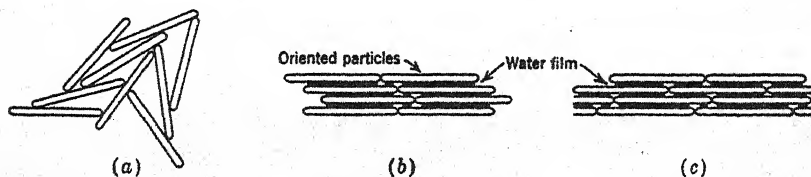


FIG. 29. Orientation of particles as a cause of soil plasticity.

the water films between the oriented plate-like colloidal particles, which impart to the soil its cohesive properties, enable the soil to be permanently molded into any desired shape or form. This moisture range corresponds to the range of plasticity of a soil. Orientation of particles and their subsequent sliding over each other takes place when sufficient water has been added to provide a film around each particle. The amount of water necessary to produce these films corresponds to the moisture content at which the soil ceases to be friable. With an excess of water, the water films become so thick that the cohesion between particles decreases and the soil mass becomes viscous and flows.

A possible mechanism of these plastic effects is shown in Figure 29 (5). The colloidal particles in a soil at low moisture contents are probably arranged haphazardly (Figure 29a). The particles become covered by an adsorbed water film as the percentage of moisture in the system increases. When a moisture content corresponding to that of the lower limit of plastic consistency is reached, the particles are

oriented when pressure is applied (Figure 29b). The tension of these adsorbed water films holds the adjacent oriented particles together. As the pressure is increased above that of the tension of these films, the particles slide over each other. After the pressure is removed, the particles do not return to their original positions because they are held in place in their new positions by the tension of the moisture films (Figure 29c). This concept permits a fairly simple explanation of the phenomenon of plasticity in soils. It takes into account the physical behavior of the soil colloids and the importance of film tension in plasticity effects. It will be noted in subsequent paragraphs that experimental evidence on the plasticity of various soils, as well as investigations in soil dynamics and tillage, tend to substantiate the film theory of soil plasticity.

Methods for Determining Soil Plasticity. Various indirect methods have been used for evaluating soil plasticity. These include the calculation of plasticity from either the solidity or the rate of slaking of dried briquettes, the use of water adsorption and the colloid content as indices of plasticity. It has been shown, however, that plasticity cannot be characterized sufficiently by indirect means, even though good correlations may exist between the aforementioned measurements and the plastic behavior of soils. Consequently, studies of soils in the plastic state have undoubtedly produced the best results.

One of the most widely used techniques for studying plasticity is the flow of clay pastes through capillary tubes. This technique is rather commonly employed in the ceramic industry. Blair and his coworkers (7, 8) have investigated the type of flow that occurs when a clay paste is forced through a capillary tube and have found that there are four stages in the volume of flow-pressure curve. These stages are indicated by curve *OABC* in Figure 26.

First stage—As pressure is applied, no flow is produced until a certain critical stress is reached. Blair considers this "yield value" as the stress necessary to overcome the adhesion of the water envelope (that separates the paste from the wall) on the wall of the tube.

Second stage—After the critical pressure to produce the initial flow has been reached, the paste moves as a solid plug along the tube. It slides through a water envelope of constant thickness. The volume of flow is proportional to the applied pressure in this region.

Third stage—Uniform flow of the plug changes to a telescopic type as the shearing stress is increased. The diameter of the central plug decreases as it becomes surrounded by a cylindrical shell

whose thickness increases with the applied pressure. The volume of flow increases more rapidly than the pressure and the curve slopes steeply upward.

Fourth stage—Finally, the central plug disappears and flow is proportional to the applied pressure as the paste moves as a liquid. If this portion of the curve, *BC*, is extrapolated to the point *D*, the so-called yield value of Bingham is obtained. This represents the shearing strength of the material. Keen and Blair have called this value the static rigidity of the paste. It represents the energy that is required to produce flow and is considered a measure of the cohesive properties of the system.

Blair has suggested the term "flow plasticity" to characterize the plasticity of different clays. It so happens that the critical pressures responsible for the type of flow in the second and third stages depend upon the radius (*R*) and length (*L*) of the capillary tubes employed in the measurements. If *a* and *b* are these critical pressures, respectively, then:

$$A = \frac{aR}{2L} \quad \text{and} \quad B = \frac{bR}{2L}$$

where *A* and *B* represent stresses per unit area on the wall of the tube. Blair states that "if pastes of clays in water are compared at such (different) concentrations of water that they show the same *A*-value, then the *B*-values measured at these concentrations are very closely related to the plasticity of the material." Flow plasticity is then defined as the value of *B* in dynes per square millimeter for a paste at such a water content that it gives an *A*-value of 1 dyne per sq. mm.

Bodman and Tamachi (9), by considering the law of plastic flow as given in equation 22, have devised a simple method to compare the moisture contents of soils at points of equal stiffness; that is, at that consistency at which the application of the same force causes the same amount of flow in various soils. Their method involves the placing of the plastic soil mass in a flat-bottomed container, making a ditch in the mass and dropping the container from a fixed height to a flat surface. When the number of impacts needed to fill the ditch are plotted as a function of the moisture content, a hyperbolic curve is obtained. A close correlation exists between the moisture content at points of equal stiffness and the colloidal-clay content. This technique gives results that are probably similar in character to the *A*-values in flow-plasticity measurements.

Atterberg (1, 2) studied plasticity from the point of view of the

moisture range over which plasticity was manifested. He suggested three values that have attained rather wide usage among soil investigators. These are (1) the upper plastic limit, or that moisture content at which the soil will barely flow under an applied force, (2) the lower plastic limit, or that moisture content at which the soil can barely be rolled out into a wire, and (3) the plasticity number, or the difference between the upper and lower plastic limits. The latter is taken as an index of plasticity.

The upper plastic limit was determined originally by Atterberg by placing a small amount of soil in a round-bottomed dish, working it into a stiff paste, pressing it tightly against the bottom, cutting a V-shaped groove in the plastic mass and jarring the dish to make the two segments flow together. If flow was not produced, additional water was added and the process repeated. If too much flow was obtained, dry soil was mixed with the plastic mass. This cut-and-try process was repeated until the correct flow was obtained. The moisture content of the plastic soil was then determined.

It is obvious that such a determination is susceptible to personal errors and is difficult to standardize so that various individuals can reproduce the same results. In light of this fact, the principle of the solution balance has been adapted to this determination with excellent results (6); Casagrande (10) perfected a special apparatus for the same purpose. Both techniques reduce the personal element to an insignificant minimum.

The lower plastic limit is determined now in the same manner as originally. Dry soil is mixed with water in a round-bottomed dish until it begins to lose its crumbly feel and shows a tendency to become plastic. The mass is then kneaded in the hands. A small portion is rolled between the fingers and a glass plate, or piece of glazed paper, until a wire is formed. The process of adding water or soil is repeated until that moisture content is reached when the plastic mass will just barely roll out into a wire that breaks into pieces about $1/4$ to $3/8$ inches long. Although this technique seems rather crude, it is possible to duplicate results with considerable accuracy. Tough clays are the most difficult to study.

Significance of the Atterberg Limits. The lower plastic limit represents the moisture content of the change from the friable to the plastic consistency. Orientation of particles and their subsequent sliding over each other take place at this point, since sufficient water has been added to provide a film around each particle. Experimental evidence shows that cohesion is a maximum at the lower plastic limit. This is illustrated in Figure 65, Chapter IX. The moisture content

of this limit depends upon the amount and nature of the colloidal material present. The colloid content regulates the number of films and the nature of the colloid determines the quantity of water that is adsorbed before a distinct water film around each point of contact is formed. This is substantiated by the fact that a high degree of correlation exists between the hygroscopicity over 30 per cent H_2SO_4 and the moisture content of the lower plastic limit. As stated before, the lower plastic limit represents the minimum moisture percentage of a soil at which it can be puddled.

The upper plastic limit signifies that moisture content at which the water films become so thick that cohesion is decreased and the soil mass flows under an applied force. It also depends upon the number of films that are present; essentially, it is the moisture content at which most of the films coalesce to fill up the majority of the soil-pore space. A close linear relationship exists between the moisture content of the upper plastic limit and the moisture equivalent, within a wide range of soils. Such a relationship suggests that a rather thick film exists around the particles at this limit. The upper plastic limit is similar in character to the A -value of Blair and the point of equal stiffness of Bodman and Tamachi. It is defined as the moisture content at which a soil will just barely flow under the applied force.

The plasticity number is an indirect measure of the film tension, that is, of the force required to mold the soil. It is entirely a function of the number of films and represents the amount of water that must be added to a soil system to increase the value of d in equation 26 from a film thickness at which maximum tension occurs to a thickness at which flow is produced. In other words, it is a measure of d from that moisture content at which the soil ceases to be friable to that where the soil begins to flow. On the basis of this equation, the plasticity number should be proportional to the ratio $F_1 - F_2/F_2F_1$, where F_1 and F_2 are the tensions at the upper and lower plastic limits, respectively. This relationship holds with widely different clay soils that have been diluted with silt to vary the plasticity number. The nature of the colloidal material changes the proportionality constant; this should be expected since the force of adhesion of water onto the surface varies with the type of colloid.

Factors Affecting the Atterberg Constants—CLAY CONTENT. Since plasticity is a function of the finer soil fractions, various soils will possess different plasticities according to the amount of clay they contain. Atterberg (1, 2) and Terzaghi (27) have shown that an increase in the percentage of clay causes both plastic limits to be higher on the moisture scale and increases the plasticity number. The results

in Figure 30a point out the same effects of clay. The clay content of the Lufkin soil was varied by additions of silt. These results are typical of a large number of samples. It is seen that the moisture content of the lower plastic limit becomes slightly less as the clay content is decreased. This should be expected since it takes more water to satisfy the surface forces of clay than of silt. The outstanding effect of a decreased clay content is the rapid lowering of the upper plastic limit and the consequent decrease in the plasticity number. It takes a considerable quantity of water to cause the large number of films in a soil containing 87 per cent of 5μ clay to become so thick that the plastic mass can be made to flow. Much less water is required when only 37 per cent of clay is present.

It was previously suggested that the plasticity number depends entirely upon the number of films. The latter is necessarily a function of the clay content. Russell (25) has reported that the plasticity number is a linear function of the clay content (5μ particles). When the results of Russell are analyzed along with data on synthetic soils, an excellent linear relationship is observed between plasticity number and the content of 5μ particles. This relationship is expressed as follows:

$$P.N. (\text{plasticity number}) = 0.6 C (\text{clay content}) - 12$$

Plasticity is not exhibited by soils that contain less than 20 per cent of 5μ particles. The results from synthetic soils show a similar relationship for 1μ particles, namely, $P.N. = 0.66 C - 10$. Soils containing less than 15 per cent clay are generally non-plastic. Any marked variations in particle shape will affect these correlations. However, in light of the wide variety of soils that are contained in the samples from which the correlations were made, these expressions clearly show the general relationship that exists in most soils between clay content and plasticity number.

NATURE OF SOIL MINERALS. Atterberg, in his original investigations on soil plasticity, was interested in finding out to what extent the different minerals, from which soils are derived, affect plasticity. Consequently, he ground various minerals to give particles the size of clay and measured their plasticities. The results of this experiment are rather outstanding in light of recent knowledge on the nature of clay minerals. His studies showed that only those minerals that have a platy or sheet-like structure exhibit plasticity when ground. Quartz and feldspar, whose crystals are made up of linked tetrahedra, are non-plastic. On the other hand, kaolinite, talc, muscovite, biotite and others, whose crystal lattices are built up in sheets, are plastic. These

differences are attributed to a greater surface and increased contact in the case of plate-shaped particles. Although it is realized that few soils contain sufficient amounts of these primary minerals to affect plasticity very markedly, nevertheless, the fact that the secondary clay minerals have sheet-like structures similar to the aforementioned plastic primary minerals helps to explain the plasticity of clays. In some cases, the presence of appreciable quantities of mica in the silt fraction may cause higher plasticities than would be expected from the clay content. It is interesting how this small portion of a rather comprehensive investigation on plasticity fits into the modern picture of crystal-

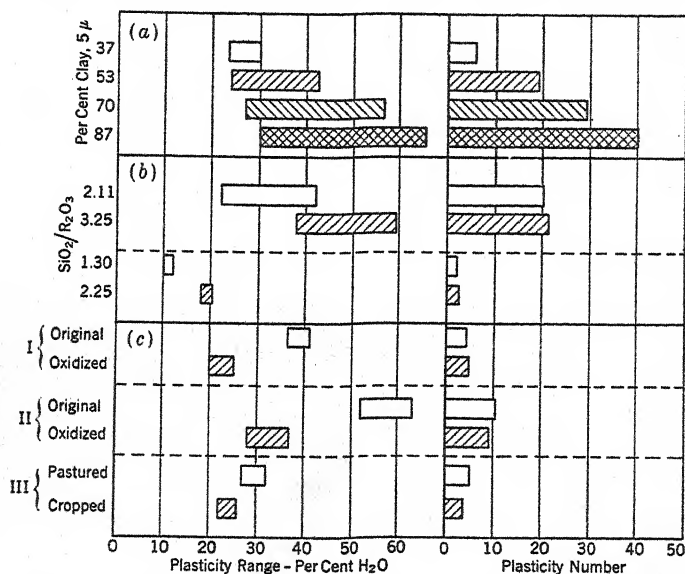


FIG. 30. Factors affecting the Atterberg consistency constants.

lattice constitution and particle shape in relation to film tension and cohesion.

CHEMICAL COMPOSITION OF THE COLLOID. It has been shown that the physicochemical properties of colloidal clay vary with the SiO₂-sesquioxide ratio. The adsorptive capacity of the colloidal surface for cations and water molecules decreases as the ratio of silica to the sesquioxides becomes lower. This same effect is manifested in the Atterberg consistency constants (5). The results in Figure 30b point out that the location of the plastic range on the moisture scale is greatly lowered by a small SiO₂-sesquioxide ratio, when two soils with the same clay content, but of different chemical composition, are com-

pared. Even though both plastic limits are lowered, the plasticity number is not greatly affected. This should be expected on the basis of the film theory of plasticity. Since moisture adsorption and, consequently, the amount of water required to produce a film around the particles is small in soils having a low ratio, the moisture content at which plasticity develops should also be low. Once a sufficient film is formed to cause plastic effects, the amount of water necessary to thicken the films to the point of flow is dependent upon the number of films. This number is the same for both types of soils. These data indicate that kaolinitic minerals become plastic at much lower moisture contents than the montmorillonitic types. It should be kept in mind that this reasoning may break down in true laterites where large quantities of the oxides of alumina and iron are present.

NATURE OF EXCHANGEABLE CATIONS. The nature of the exchangeable cation has considerable influence upon soil plasticity (3). In general, the Na and K ions produce similar effects; Ca and H ions affect plasticity similarly. The K ion causes a lowering of both plastic limits and a decrease in the plasticity number. The Na ion lowers both plastic limits but increases the plasticity number. The H ion tends to increase the plasticity of those soils which are originally in a state of unsaturation. The Mg and Ca ions increase plasticity, both plastic limits being lowered. These variations perhaps can be explained on the basis of the differences in the hydration of these clays.

There are two types of hydration affecting the plasticity values, namely, water of hydration held between the particles in aggregate formation and the hydration hull or film around the individual particles. In order to produce plastic flow there should be an excess of water present above that required to satisfy the sorptive capacity of the particles or aggregates for water. The K-saturated soils require less water to produce these plasticity effects than the Na-saturated soils. This is merely a difference in the hydration of the particles as a function of the hydration of ions. The Ca- and Mg-soils require a certain amount of water to fill the pores in their aggregates; therefore, a larger quantity of water is necessary to produce plasticity than in the case of the K-saturated soils. The high hydration and dispersive action of the Na ion make the plasticity of the Na-saturated soil greater than those soils saturated with the divalent ions. The magnitude of the effects of the adsorbed ion varies with the amount and nature of the clay.

ORGANIC-MATTER CONTENT. Organic matter exerts an interesting effect upon soil plasticity. Measurements of the plasticity constants of different soil profiles usually show that the plastic limits of the surface

horizon are fairly high on the moisture scale, when compared with the other layers. This effect is apparently associated with the presence of organic matter in the surface horizon. Oxidation of the organic matter with hydrogen peroxide causes a decided lowering of both plastic limits (4). This is illustrated in Figure 30c. Soil I, with an organic matter content of 3.5 per cent, became plastic at 36.5 per cent moisture. Removal of the organic matter lowered this limit to 19.8 per cent moisture. Moreover, the oxidized soil flowed at 25.1 per cent moisture, whereas the soil with the organic matter was friable up to a moisture content of 36.5 per cent. The plasticity number was not materially changed by oxidation of the organic matter. Soil II, which contained 7 per cent organic matter, became plastic at 52.2 per cent moisture. Removal of the organic matter caused the soil to exhibit plasticity at 27.7 per cent moisture.

Plasticity measurements on the virgin and cultivated Putnam silt loam (Soil III) in Missouri show the same effects of organic matter. The virgin area contained 3.9 per cent organic matter and the cultivated 2.6 per cent. This decrease in organic matter has taken place during 60 years of cultivation. The lower plastic limit of the virgin field was 27 per cent moisture; that of the cultivated was 22 per cent. Thus, oxidation of organic matter in the field through natural means produced results similar to artificial oxidation in the laboratory.

The causes of this decided lowering of the plastic limits on the moisture scale, without a really significant effect upon the plasticity number, are readily understood on the basis of the film theory of plasticity. Organic matter has a high absorptive capacity for water. Hydration of the organic matter must be fairly complete before sufficient water is available for film formation around the mineral particles. Consequently, the lower plastic limit occurs at relatively high moisture contents. After the films are formed, however, practically all the additional moisture functions only to enlarge the films until flow is produced. The presence of organic matter has little effect on this type of water and, therefore, does not influence the plasticity number to any significant extent.

The results in Figures 30b and 30c point out the importance of considering the range of plasticity on the moisture scale in discussing soil consistency. It is evident from these data that two soils may possess the same plasticity number but exhibit plasticity at entirely different moisture contents. The practical significance of the plasticity range is well illustrated by Soil II, Figure 30c. The presence of organic matter makes it possible to cultivate up to 52.2 per cent moisture without puddling the soil. When the organic matter is removed, the soil pud-

dles at 27.7 per cent moisture. Thus, the addition of organic matter to soils may be expected to extend the zone of friability to fairly high moisture contents.

Practical Significance of the Atterberg Constants. The original work of Atterberg was apparently conducted with the hope of obtaining some physical criterion for the classification of soils. He made numerous consistency determinations in an attempt to classify the soils of Sweden. Terzaghi (27) has also suggested that the plasticity limits may serve as an index for the physical classification of soils. Soils with a high upper plastic limit should either contain a large quantity of the excessively fine-grained fractions or be rich in plate-like particles. Soils having a high upper plastic limit and a low plasticity number should be in a finely divided state. If the plasticity number is high, they should contain an abundance of plate-like particles.

Russell (25) has found that the Atterberg constants are a very satisfactory index of soil consistency and of the degree of clay accumulation in the soil profile. So far as soil classification is concerned, these constants will serve as an index of the accumulation of clay within any given soil profile and express the differences in the consistencies of various soils. The question remains, however, whether they can be used to classify different soils in light of the various factors that affect their value.

Terzaghi (27) has related the lower plastic limit to the permeability of clays and to the rate of evaporation of water from soils. He has stated that the coefficient of permeability of a homogeneous clay decreases rapidly with decreasing water content until, at the lower plastic limit, it becomes practically equal to zero, regardless of the value of the plastic limit. He also has stated that the rate at which water evaporates from the surface of a clay sample is about 4 per cent greater than the speed with which it evaporates from a free water surface, providing the moisture content is higher than the lower plastic limit. For moisture contents equal to or smaller than the lower limit of plasticity, the physical properties of the soil water are no longer identical with those of free water.

Terzaghi also has reported that the compression of a soil increases rapidly with moisture above the moisture content of the lower plastic limit. The compression is small for low moisture contents. The break in the curve, showing the relation of the deformation of a clay to the percentage of moisture, corresponds approximately to the lower plastic limit. This break is called "the critical bearing point."

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CHAPTER V

SOIL STRUCTURE

DEFINITION

Soil structure is usually defined as the arrangement of the soil particles. This concept, however, requires a clear understanding of the word, "particles." As far as structure is concerned, soil particles refer not only to the individual mechanical elements, such as sand, silt and clay, but also to the aggregates or structural elements, which have been formed by the aggregation of smaller mechanical fractions. The word, "particle," therefore, refers to any unit which is a part of the makeup of the soil, irrespective of its being a primary (sand, silt or clay fraction) or a secondary (aggregate) particle. Consequently, the structure of a soil implies an arrangement of these primary and secondary particles into a certain structural pattern.

Zakharov (53) has placed a slightly different interpretation on structure, which he defines as "the very fragments or clods, into which the soil breaks up." The capacity of a soil to break up into fragments or aggregates of different sizes and shapes is referred to as its structure capacity.

The structure of any particular horizon of a soil profile as it appears to the eye of the observer may be considered as the macro-structure of the soil in this layer. That is, a more or less definite picture of the arrangement of the particles is distinctly evident. This visible macro-structure, however, is undoubtedly dependent upon the type of arrangement of those soil particles which are not so clearly distinguishable to the naked eye. In other words, the nature of the arrangement of the smaller units, or micro-structure, markedly influences the macro-structure.

CLASSIFICATION OF STRUCTURE ✓

Classifications Based upon the Size, Shape and Character of the Aggregates. Zakharov has suggested a classification of soil structure on the basis of the size, shape and character of the surface of the soil aggregates, fragments and clods. The type of structure is distinguished by the main shape of the structural particles. Three

principal types are recognized. They are (1) cube-like structure, in which the secondary particles are equally developed along the three axes, (2) prism-like structure, in which the secondary particles are elongated in the direction of the vertical axis, and (3) plate-like structure, in which the structural units are shorter in the vertical direction and are developed more in the direction of both horizontal axes.

The kind of structure within each of these principal types is distinguished on the basis of the character of the faces and edges of the aggregates. Two kinds of structural units are recognized, namely, those in which the faces and edges are indistinctly manifested and those in which the faces and edges are clearly defined. For example, the main difference between cloddy and nutty structure is simply that the former has indistinct faces and edges. The fact that structural units generally break out along certain planes of cleavage adds emphasis to the use of the character of the surface as an important criterion for classification purposes. Acute, oblique or rounded edges are usually formed when cleavage takes place. The resulting faces may be smooth, shiny and dense, or rough and somewhat open.

Finally, Zakharov distinguishes the structural species on the basis of size. Coarsely and finely granular structures are two species within the granular kind of structure in the cube-like type. Zakharov's classification of structure is presented in Table 13.

A modified version of the Zakharov classification was suggested by the Committee on Soil Structure of the American Soil Survey Association (41) in 1928. The proposed classification divided soil structure into two major groups. They were simple structure, in which the particles are not flocculated or aggregated, and compound structure, in which particles are flocculated into aggregates or granules. A summary of this classification is presented in Table 14. It is important to note that, in addition to making several differentiations within the simple-structured soils, there is a subdivision of granular and fragmental kinds of structure within the cube-like type. In other words, this classification infers that certain soil aggregates may represent fragments resulting from a breaking down of larger units. The committee, in discussing granular and fragmental structure, has stated that "it is believed that there is a very real difference between these, a difference that is possibly based on processes of aggregation that can only be conjectured at the present state of knowledge."

A modification of the Zakharov classification, on the basis of the type of arrangement of the particles within the structural unit, has been suggested (4). As a result of a large number of determinations

and observations on the amount and character of the secondary particles from numerous soils, the micro-structure of soils is divided into three main groups or classes. These groups are differentiated by the nature of the arrangement of the soil particles, as shown in Figure 31. The structural arrangement of a soil may consist of (1) chiefly secondary particles, (2) chiefly primary particles or (3) a mixture of secondary and primary particles, in which there tends to be single-grained arrangement within the secondary units. In studying an arrangement due chiefly to secondary particles, it was observed that the relatively large secondary units are made up of smaller secondary ones. Such units are almost always cube-like or spherical, have more or less rounded edges, are distinctly pervious and break up not along well-defined cleavage planes but by a complete crumbling of the entire secondary particle into smaller ones, with properties similar to the larger. These secondary particles are true granules, and the soil has a granular structure. Granulation refers to the process of formation of this type of secondary particles. The soil mass crumbles into the numerous granules when it is handled.

The secondary units in certain soils, which have clearly defined angular edges, are distinctly different from these more or less rounded and pervious granules. These angular particles are generally dense and not very pervious. They break up with the formation of smaller units, similar in shape and consistency. The breaking up takes place along well-defined cleavage planes or surfaces of fracture. Inasmuch as these secondary particles originate as a result of fracturation along definite faces, and since the units thus formed are angular and non-pervious, the term fragmentation is suggested to describe this breaking down process. The secondary units formed as a result of fragmentation are classified into two subdivisions, the prism-like and plate-like secondary particles. The arrangement within the secondary units consists almost wholly of primary particles. This single-grained arrangement within the secondary units is largely responsible for the density, angularity and shiny faces of the fragments.

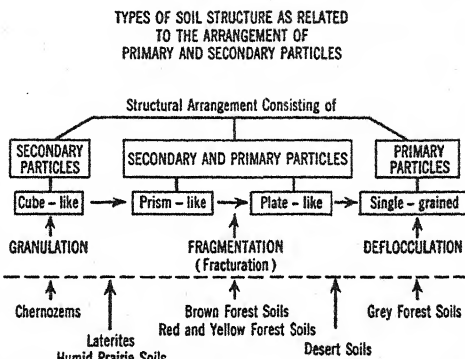


Fig. 31. Classification of soil structure according to Bayer (4).

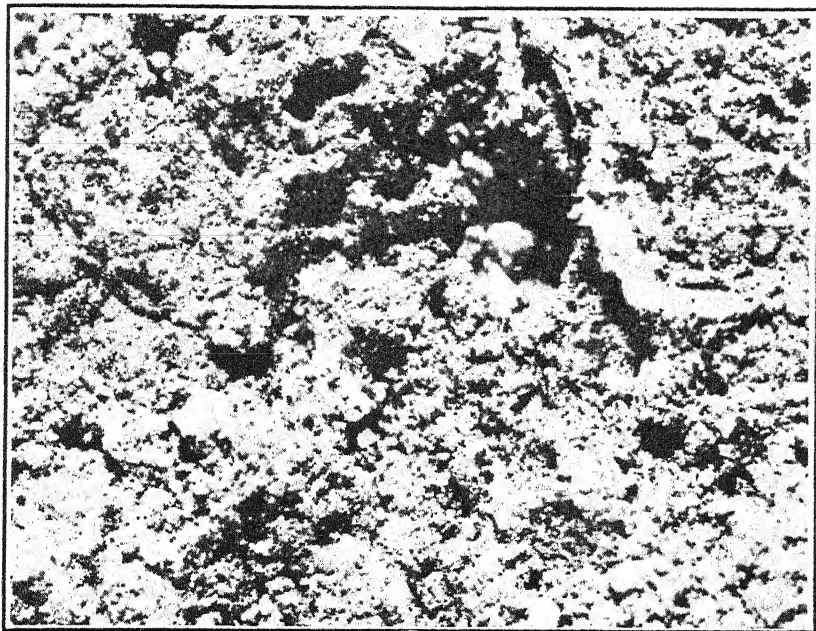
The structural arrangement of soils consisting of primary particles is self-evident. It is the predominating structure in the coarser-textured soils. Since no secondary units are present, the single-grained arrangement of the primary particles makes the soil appear in a state of deflocculation.

The differences between these types of units are clearly shown in the photomicrographs in Plate 1. These pictures represent the surface of a structural unit about the size of shellbark hickory nut; the magnification is about 20 times. Picture *A* portrays the nature of the surface of a unit from a chernozem-like soil from Canada. It is seen that this unit consists of a large number of small, rounded granules that seem to be made up of smaller granules of the same shape. This arrangement gives an impression to the eye of softness and perviousness.

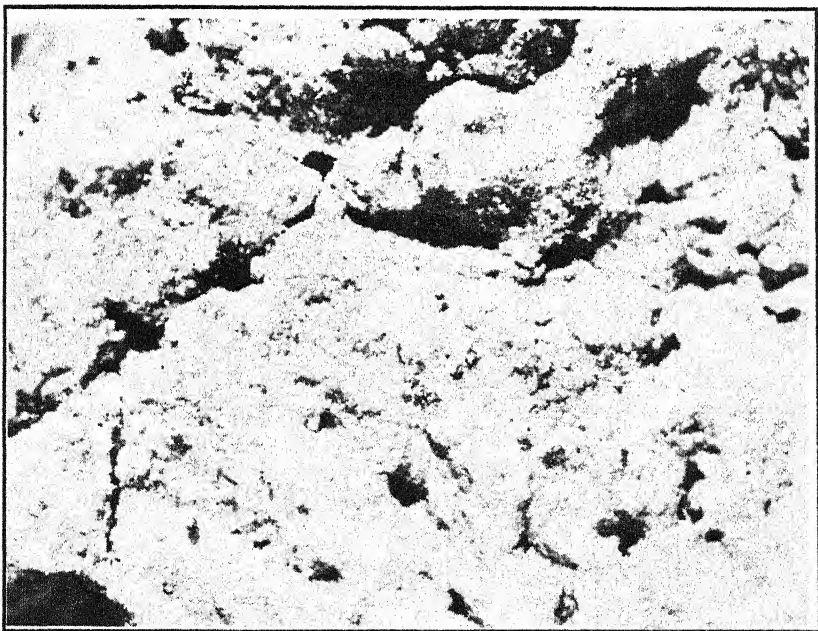
Picture *B*, on the other hand, which represents the surface of a unit of equal size from the B-horizon of a prairie soil, gives a very different impression. It is observed that the unit is dense, compact and impervious. The angular shape of this type of unit is shown by the small fragment in the lower left-hand corner, which is breaking off along a definite cleavage plane. There is a distinct single-grained arrangement of the particles within this secondary unit. The formation of secondary particles in this type of soil is indeed a process of fragmentation.

These pictures, along with numerous laboratory determinations on aggregation and porosity, suggest the desirability of including the internal makeup of the aggregate in a classification of structure. The word "granular," for example, is perhaps one of the most loosely used terms in discussions of soil structure. It is employed to define the type of pervious and rounded aggregates that are so characteristic of the A-horizon of chernozems. It is often used to designate the dense, angular aggregates that are present in many subsoils. Fortunately, however, there is a trend among American soil scientists to recognize at least the limitations in the usage of "granular." At the moment, the term "fragmental" is being suggested to characterize the dense, angular type of aggregate. Although this terminology may not prove to be adequate, nevertheless, it seems to be a step in the right direction.

Other Classifications. In addition to the classifications of soil structure that are based upon the size, shape and character of the aggregates, several groupings have been suggested that are based upon the type of pore space and the nature of the binding material that is responsible for aggregate formation. For example, the use of the nature of the pore space for classification purposes has been suggested



A



B

PLATE 1. Photomicrographs of granular (*A*) and fragmental (*B*) types of structure (magnification about 20 times).

by Pigulevsky (36), who impregnates the soil with wax, examines a thin section under the microscope and differentiates various structures by the type of cracks or pores observed.

Kubiena (29) has proposed a classification of micro-structure that is based upon the nature of the "soil fabric" (arrangement of the constituents of the soil in relation to each other). This is a summary of several of the essential features of his tentative groupings:

COATING OF MINERAL GRAINS	ARRANGEMENT OF THE "FABRIC"	OCCURRENCE
Grains not coated	1. Grains embedded loosely in a dense ground mass	Lateritic soils
	2. Grains united by intergranular braces	Chernozems, brown earths, lateritic soils
	3. Intergranular spaces containing loose deposits of flocculent material	Sandy prairie soils
Grains coated	1. Grains cemented in a dense ground mass	Desert crusts Podsol B-horizons
	2. Grains united by intergranular braces	B-horizon of podsolized, brown forest soils
	3. Intergranular spaces empty	B-horizon of iron and humus podsols

This grouping has resulted from microscopic examination of soils. It is interesting to note that there is no colloidal coating on the mineral grains in those soils which are usually considered to have a granular-like structure. It is possible that the type of micro-aggregate will play an important part in helping to set up future classifications of structure, especially for detailed separations. It seems logical to suggest, however, that the readily visible aspects of soil structure are more technically and practically useful as means for classification in a broad sense.

In considering the relation of the various types of structure to each other, it seems that the degradation of a granular structure leads to the formation of the fragmental type and thence to the single-grained state. On the other hand, the formation of secondary units in a soil, in which single-grained structure predominates, appears to take place with the production of plate-like units.

These types of structure can be correlated fairly well with the typical soils within each of the main soil groups. This is suggested in Figure 31. True granulation, when the structural arrangement consists of secondary particles, is best expressed in the chernozem-like soils, particularly the surface horizon. This type of structure tends to

change with increasing depth towards fragmentation. Podsol exhibit a single-grained arrangement in the upper layers which grades into a platy type with depth, and thence into a fragmental grouping.

The proposed relationships that are suggested in Figure 31 are the result of aggregation studies on a large number of soils. They confirm, in part, the views of Zakharov, who has visualized the structure of the various soil groups as follows: "The upper horizon of a typical loamy chernozem is of a clearly exhibited powder-like and granular structure, owing to the high content of humates, while the lower horizon assumes a nutty structure, which is deeper on replaced by prism-like structural aggregates. The podzolized soils, in their upper horizons, bear a feebly expressed cloddy-pulverescent structure altering under the influence of a higher content of silicic acid in the underlying layer into a platy-pulverescent one; deeper the structure becomes platy, then flat-nutty, and still deeper acquires a prismatic character. In the upper horizon of a brown alkali-soil, we frequently meet with a squamose or foliated structure which, with increasing depth, is suddenly replaced by clearly shaped columnar fragments with rounded upper bases ('tops'), and changes further into a sharp-faced structure."

Granular structure is apparently at a maximum in the surface of the chernozem and chernozem-like soils. It gradually becomes less pronounced as one passes through the humid prairie to the forest soils or through the chestnut to the desert soils. The surface of the gray forest soils, or podsol, and of the desert soils has single-grained to plate-like structures. Lateritic soils occupy an intermediate position between granulation and fragmentation.

The B-horizon of the chernozem-like soils has a prismatic structure; that of most brown, red and yellow forest soils is fragmental; and that of the true podsol is seemingly a transition between the platy and fragmental types. These differences are related to the structure-forming factors that vary with climate; consequently, variations in the morphology of the profile exist.

THE GENESIS OF SOIL STRUCTURE

The genesis of soil structure refers to the causes and methods of formation of the structural units or aggregates. It has already been shown that an aggregate consists of an intimate grouping of a number of primary particles into a secondary unit. The mechanism of the formation of these aggregates is one of the most important phases of the soil-structure problem.

Since many of the present-day concepts of granulation and aggre-

gate formation are built on *a priori* reasoning from the phenomenon of flocculation in dilute suspension, it is necessary at this time to distinguish between flocculation from the purely colloidal point of view and aggregation from the standpoint of soil structure. *Flocculation and stable-aggregate formation are not synonymous. The former is primarily electrokinetic in nature. Primary particles with a high electrokinetic (zeta) potential repel each other when they collide in a suspension. When the potential is lowered sufficiently, a collision between particles results in a mutual attraction and the formation of a floccule. The floccule is stable as long as the flocculating agent is present.

*Stable-aggregate formation requires that the primary particles be so firmly held together that they do not disperse in water. In other words, from the point of view of soil structure, aggregate formation requires a cementation or binding together of flocculated particles. Thus, flocculation may aid in the aggregation process, but is not aggregation in itself. As Bradfield (9) so aptly states, "Granulation is flocculation plus!" It is necessary to keep this fact in mind to avoid erroneous interpretations of flocculation data in terms of stable-aggregate formation.

Cation Effects in Aggregate Formation. It is almost universally recognized that lime and organic matter improve the physical properties of the soil. Moreover, it has been demonstrated quite conclusively that the poor structural qualities of alkali soils can be changed into a favorable physical condition if the sodium is replaced by calcium. These facts, along with the laboratory observations that clay suspensions can be flocculated by calcium salts, have led to the widely accepted viewpoint that the beneficial effects of lime are due to its ability to flocculate the soil colloids. The influence of organic matter has been attributed to its cementation effects.

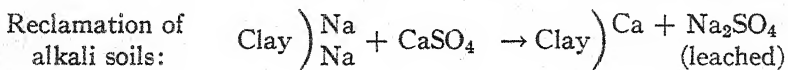
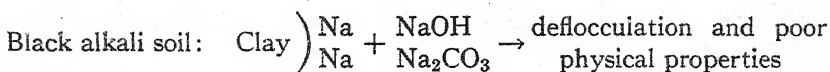
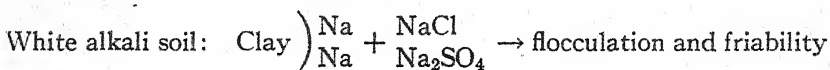
There is little doubt that the effect of calcium on the reclamation of alkali soils is primarily associated with the more favorable influences of this ion on the properties of the soil colloids, as compared with the effect of the sodium ion. Sufficient evidence has been presented previously to show that Na-saturated soils are more highly hydrated and dispersed than Ca-saturated soils. (See Table 10.) The former swell and become impervious; this is not true of Ca-soils. The results in Figure 32 show the differences in permeability of Na- and Ca-clays. Consequently, the reclamation of alkali soils is based upon the removal of sodium ions from the base-exchange complex by replacement with calcium ions. The replaced sodium must then be removed from the soil through leaching. Irrigation waters must not contain appreciable amounts of sodium, but ample quantities of calcium salts should be

present. Thus, the flocculating ability of calcium is the main factor in the betterment of alkali soils.

The properties of different alkali soils vary, dependent upon flocculation effects that are produced by the presence of excess salts. Even though an abundance of sodium ions is characteristic of all alkali soils, the influence of sodium on the physical properties depends upon the nature of the sodium compounds that are present. For example, white alkali soils contain sodium on the exchange complex and an excess of sodium salts (NaCl , Na_2SO_4) in solution. These soils are fairly friable and are not highly dispersed. The reason for this friability is found in the flocculating effects of the excess salts. The presence of sodium salts causes a repression of the "ionization" (thickness of the electrical double layer) of the adsorbed sodium ions. This causes a decrease in potential, hydration, swelling and dispersity. (See Figure 20.) The colloidal material is flocculated.

Black alkali soils, however, present a different picture. Rainfall has just been sufficient to leach out the excess salts and the soils are characterized by a Na-saturated exchange complex. The Na-saturated colloidal material hydrolyzes to form NaOH and Na_2CO_3 . Such a system is highly dispersed, highly hydrated and impervious to water. Its state of deflocculation is responsible for the poor physical properties. The dark color is due to the solubility of small amounts of organic matter because of the Na_2CO_3 .

The relation of the adsorbed ions to alkali-soil problems may be briefly summarized by the following schematic reactions:



A different situation exists when the effects of calcium on the aggregation of acid soils are thoroughly analyzed. It has been generally accepted in the past that the flocculating effect of the calcium ion is the contributing factor for stable granulation. Recent experimental observations, however, indicate that the direct effect of the calcium ion on the aggregation of acid soils is not as important as was originally considered. Early experiments on the effect of exchangeable cations on the physical properties of soils (1) indicated that H- and Ca-saturated soils are somewhat similar. In some cases,

H-soils were more flocculated than Ca-soils; in other instances, the reverse was true. Later studies (2), on the effect of calcium saturation on the properties of a colloidal clay, distinctly showed that calcium had to be present in excess of the saturation capacity to bring about better flocculation than the hydrogen ion. Lutz (30) then reported that Ca- and H-saturated clays do not differ much with respect to swelling and permeability. In all instances, however, H-clays were slightly more permeable than the corresponding Ca-systems. This fact is illustrated in Figure 32. These curves represent the permeability of colloidal-clay membranes for water.

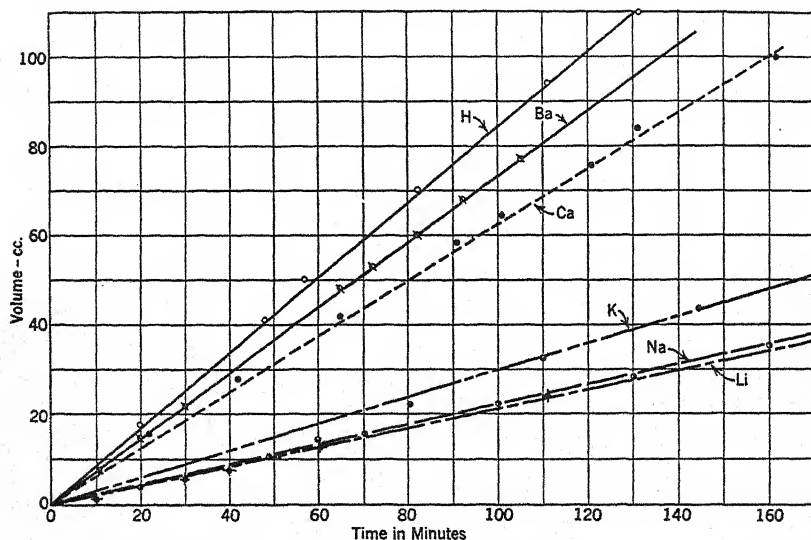


FIG. 32. The effect of the nature of the adsorbed ion on the permeability of clays (Lutz, 30).

A statistical analysis of 77 different soils of the United States showed that there is no significant correlation between the amount of exchangeable calcium and granulation (4). This lack of correlation is explained by the data in Table 15. These results were obtained by first leaching the soils with either 0.01 *N* HCl or 0.1 *N* CaCl₂ and then with water, until free from chlorides, so as to give H- and Ca-saturated systems. It is evident that there is no significant difference between these two systems. Replacement of calcium ions by hydrogen did not cause dispersion.

Bradfield (9) has reported poor structural conditions in two heavy soils from Ohio. The one, a heavy glacial lake clay, was completely

TABLE 15

THE EFFECT OF REPLACING EXCHANGEABLE CALCIUM IONS BY HYDROGEN IN THE AGGREGATION OF CHERNOZEM-LIKE SOILS

Soil type	Carbon, per cent	Exchange capacity, m.e./100 gm.	Aggregates > 0.02 mm.	
			H-saturated soil, per cent	Ca-saturated soil, per cent
Deep park fine sandy loam..	5.91	31.25	11.1	13.8
Bearden silty clay loam....	3.61	33.75	36.4	34.5
Barnes loam.....	3.35	33.8	17.1	15.6

flocculated in suspension and was not benefited by applications of lime. The other, a calcareous subsurface horizon, containing 6 to 8 per cent CaCO_3 , was practically impervious to water. Dried lumps slaked in water to give particles small enough to pass through a 100-mesh sieve. These two examples, along with other results, indicate that, while calcium may help to obtain good tilth in heavy soils, it does not necessarily insure good granulation.

Kappen (26) also maintains that liming does not always affect the physical properties of soils. He has concluded that "the replacement of hydrogen in the soil silicates by calcium by no means has given a cause for granulation." He is of the opinion that the beneficial effects of calcium are obtained through the influence of calcium upon organic matter.

The fact that no direct effect of calcium upon the physical properties of the inorganic soil colloidal fraction could be obtained suggested that the influence of adsorbed calcium may be associated with the organic colloidal fraction. Experimental investigations, however, showed a close similarity between the properties of Ca- and H-humates (5). Organic colloids exhibited about the same differences between the Ca- and H-systems as clay. It was observed that dried Ca-humus was more reversible than H-humus, which suggests a greater stability of H-humates in water. This fact was substantiated by Myers (35), who has reported that H-humate forms more stable aggregates with clay than Ca-saturated humus.

These results call attention to the fact that flocculation is not the same as granulation. In order to have stable granulation there must be a cementation of the flocculated particles. It is obvious that adsorbed

calcium is not a cementing agent. Most of the cementing agents in soils are the irreversible or slowly reversible inorganic and organic colloids. It seems, therefore, that the effect of calcium upon aggregate formation is indirect, that is, it affects the production and decomposition of organic matter.

✓ **Cementation Effects in Aggregate Formation.** COHESION BETWEEN COLLOIDAL-CLAY PARTICLES. The soil colloidal material is responsible for the cementation of primary particles into stable aggregates. Stable-aggregate formation cannot take place in sands or silts in the absence of colloids. The soil colloidal material may be divided into at least three distinct groups as far as its cementation effects are concerned. They are (1) clay particles themselves, (2) irreversible or slowly reversible inorganic colloids, such as the oxides of iron and alumina, and (3) organic colloids.

It should be expected that the formation of secondary particles from primary separates would be related to the amount of finer particles in the soil that may serve as material to be aggregated. There is a high degree of correlation between the amount of $5\ \mu$ clay in soils and the percentage of aggregates larger than 0.05 mm. in diameter; this correlation for a large number of soils was found to be 0.566, with 0.21 being significant (4). The correlation of aggregation with $1\ \mu$ clay (0.379) was not as significant as that with the $5\ \mu$ fraction. Moreover, the correlation between the percentage of aggregates larger than 0.1 mm. and clay was less than in the case of the smaller secondary particles. This correlation for the $5\ \mu$ and $1\ \mu$ clay fractions was 0.323 and 0.23, respectively. Apparently, the cementation effects of clay were more pronounced with the smaller aggregates. It was also observed that there was a higher correlation between clay content and aggregation as the amount of organic matter decreased. This fact is pointed out by the data in Table 16. At the higher percentages of organic matter, the effect of clay in secondary particle formation becomes insignificant. It is also interesting to note in this table that the effect of organic matter is less significant in soils containing large amounts of clay.

Not only is the total quantity of aggregates in a soil dependent upon the amount of clay, but also the extent to which the finer soil separates are aggregated into larger secondary units. That is, the percentage of clay that is present in the form of aggregates varies with the amount present. Apparently, the clay particles function as binding agents in themselves. In the aforementioned group of soils there was a correlation of 0.428 (with 0.21 being significant) between the clay content and the percentage of mechanical separates larger

TABLE 16

CORRELATION OF AGGREGATION WITH CLAY AND ORGANIC MATTER

Number of samples	Per cent carbon	Correlation between aggregates larger than 0.05 mm. and 5 μ clay
19	<1	0.947 (0.456 significant)
33	1-2	0.506 (0.347 significant)
25	>2	0.376 (0.396 significant)
	Per cent 5 μ clay	Correlation between aggregates larger than 0.05 mm. and organic matter (per cent carbon)
15	>35	0.663 (0.514 significant)
29	25-35	0.650 (0.370 significant)
33	<25	0.758 (0.30 significant)

than 0.05 mm. that were aggregated into secondary particles larger than 0.05 mm. The cohesive forces between the colloidal surfaces were undoubtedly responsible for this cementation action.

Russell (40) has suggested that aggregate formation is dependent upon an interaction between the exchangeable cations on the clay particle and the dispersion of liquid. His investigations indicate that the formation of aggregates is limited to particles smaller than 1 μ in diameter, is a property of only those clays having a relatively high base exchange capacity and is brought about only by those liquids, the molecules of which have an appreciable dipole moment.

The hardness of the aggregates depends upon the sizes of the clay particles, of the adsorbed cations and of the molecules of the liquid between the particles. Soft aggregates result from large sizes of either of these variables and hard aggregates from small sizes. The reversibility or irreversibility of the aggregates varies with the nature of the exchangeable cations and the wetting liquid.

Russell presents the following theory of the mechanism of aggregate formation:

Each particle is surrounded by an electrical double layer, the outer one being diffuse and consisting of cations, while the inner layer consists of negative charges presumably anchored on the surface of the particle. The cations in the diffuse layer move about in the water in the same way as they do around a complex anion, as pictured in the Debye-Hückel theory of strong electrolytes. Since the water molecules possess a dipole moment, they

tend to be orientated along the lines of electric force radiating from each ion in the diffuse layer and from each free charge on the surface of the clay particle. Every cation and particle is thus surrounded by an envelope of orientated water molecules, and the orientation manifests itself as an apparent adsorption or immobilization of water by the clay. Some of the water molecules near an electric charge may be so strongly orientated that they appear bound to it, their heat motion merely making them oscillate about the lines of force, while those further away from the charge possess only a statistical orientation around it. A clay particle in a dilute suspension can, therefore, be pictured as consisting of a central core surrounded by a surface carrying a negative charge. Around each negative charge is an envelope of water molecules which are more strongly orientated the nearer they lie to the charge. Outside this surface are the cations, also possessing envelopes of orientated water molecules. Some cations are so close to a negative charge on the surface of the clay particle that the two water envelopes belonging to these charges overlap and the water molecules in this region are orientated in their joint field. This orientation is very strong since the negative end of the water dipole is attracted to the cation and the positive end to the particle's surface.

As the water is removed the deflocculated clay suspension becomes more concentrated and an increasing proportion of the water molecules become orientated in the joint field of a positive and a negative charge. It is reasonable to assume that a certain number of cations will share their orientated envelopes with two clay particles. A linking system is thus set up consisting of: particle—orientated wetting molecule—cation—orientated wetting molecule—particle.

As removal of water proceeds, an increasing proportion of cations share their water envelopes with two clay particles, and so the number of links increases. The links also become stronger, because they become shorter. . . . In consequence, the cohesion of the clay particles, i.e., the hardness of the crumbs, increases.

This hypothesis is in accord with the generally accepted concept of oriented adsorption of liquid molecules on colloidal surfaces. The extent of this adsorption increases with the polarity of the liquid with which the surfaces are wetted. Moreover, if two surfaces are being held together by oriented liquid molecules the cohesive force increases the shorter the chain of connecting molecules. Also, it has been shown in previous discussions that, for a given type of surface, the energy of adsorption increases with the exchange capacity of the material; that is, the number of bonds in the surface varies with the cation exchange capacity. This does not mean, however, that the cations are the primary cause of water or other liquid adsorption. As previously mentioned, there is sufficient evidence to justify the assumption that the same bonds that attract cations are also responsible for the orienta-

tion of liquid molecules. It seems, therefore, that the chief weakness in the Russell hypothesis is the emphasis that is placed upon the cation as the connecting link between particles. The same phenomenon can be explained on the basis of the orientation of liquid molecules on colloidal surfaces as developed by Langmuir and Harkins.

Undoubtedly, liquid bonds are extremely important in the cohesion between clay particles. The cohesive forces between dried clay particles, however, are probably due to other bonds in the surface that are responsible for the tenacity with which the aggregates are held together. In addition, the hardness of the aggregate also depends upon the extent to which the particles are oriented to give maximum surface contact. The disc-shaped nature of most colloidal-clay particles is more conducive to cementation effects than the more or less cube-like or spherical shape of most artificial zeolites, even though the exchange capacity of the latter is several times higher than that of clays.

Sideri (43) has shown that clay is adsorbed on the surfaces of sand grains. After dehydration the adsorbed clay is very slowly reversible. The tenacity of the bonds between the clay and sand increases with decreasing particle size. He does not consider the tenacity to be dependent upon the nature of the adsorbed bases. The adsorption of clay onto the sand is explained upon the basis of similarities in their forms of orientation. Orientation is associated with the capillary forces of the moisture film between the particles and is not limited to only one layer of clay particles. Thus, the oriented adsorption of disc-shaped clay particles onto sand surfaces, and subsequent dehydration into an almost irreversible state, are considered the bases for stable-aggregate formation.

Henin (19) attaches considerable significance to the orientation of clay particles in secondary particle formation. A parallelism is noted between the forces responsible for orientation and cohesion. Water is considered to be the bond between the oriented particles. Desiccation or dehydration contracts the chain of water molecules and brings the particles together. Complete dehydration is visualized as effecting a union of the clay micelles through oxygen linkages, as water is driven off from the OH groupings in the surface. When this type of cementation takes place the system is very slowly reversible and the rehydration of the clay increases slowly with time. The only essential difference between this concept and Russell's is in the emphasis placed upon the role of the exchangeable cations.

Thus, it is seen that the cohesive forces between oriented clay particles are extremely important in aggregate formation. This is particularly true of those aggregates in which there is a dense and compact

arrangement of the individual particles. Dehydration is a basic requirement for the formation of secondary particles. The rate of reversibility in water depends upon the temperature of dehydration and the amount of orientation of the particles. The rehydration of aggregated particles has been thoroughly discussed in Chapter II.

ORGANIC MATTER AND AGGREGATION. It has been recognized for a long time that organic matter serves as a granulating agent in soils. It is a well-observed fact that organic matter must be made soluble or oxidized to insure complete dispersion of soils in mechanical analyses.

A correlation of 0.559 (with 0.21 being significant) has been observed (4) between the percentage of aggregates larger than 0.05 mm. and the carbon content of a large number of different soils. This correlation for aggregates larger than 0.1 mm. was 0.687, which indicates that organic matter is conducive to the formation of relatively large stable aggregates. If various soils are grouped according to their clay contents, it is seen that the effect of organic matter is more pronounced in those soils containing the smaller amounts of clay. This is illustrated in Table 16. A very high correlation exists between organic matter and aggregation in soils containing less than 25 per cent clay. For clay contents above 35 per cent the correlation is significant, but not nearly as high. The extent to which the finer mechanical separates are aggregated is also significantly correlated with the percentage of carbon. There was a correlation of 0.515 between the percentages of carbon and the mechanical separates smaller than 0.05 mm. that have been aggregated into units larger than 0.05 mm.

In comparing the relative effects of the clay and organic-matter contents of this large group of soils, both are found to be equally significant in causing granulation. The correlation between the percentage of aggregates larger than 0.05 mm. and 5 μ clay was 0.566; between aggregates and organic matter, 0.599; for aggregates larger than 0.1 mm., this correlation was 0.323 and 0.687, respectively, for clay and organic matter. A multiple correlation between the amount of aggregates larger than 0.1 mm. and the combined effects of organic matter and clay was 0.75. Thus, clay and organic colloids seem to be responsible for the major portion of the soil aggregates. Organic matter, however, appears to be more effective in causing the finer fractions to aggregate since the extent to which the silt and clay separates were aggregated showed a correlation of 0.515 with organic matter and of 0.428 with 5 μ clay. These results suggest that organic colloids not only cause a high degree of aggregation of the clay particles but that they also produce large aggregates.

It is interesting to note that, even though desert soils are poorly

aggregated, the amount of aggregates that is present is correlated with the small quantities of organic matter (6). The only group of soils in which a correlation has not been observed between organic matter and aggregation is the lateritic soils, where, as will be discussed in the next section, dehydrated oxides of alumina and iron are responsible for stable aggregate formation.

Demolon and Henin (13) have found that colloidal organic matter is more effective than clay in causing the formation of stable aggregates with sand. Myers (35) has confirmed these findings, using fine quartz sand and orthoclase particles as material to be aggregated. His results indicate that dehydration is essential to aggregate formation. In all instances H-saturated systems were more highly aggregated than the corresponding Ca-systems. A summary of Myers' data is given in Table 17. It is seen that colloidal clay causes the cementation

TABLE 17

SYNTHESIS OF AGGREGATES FROM SAND * AND ORTHOCLASE BY ORGANIC COLLOIDS (MYERS, 35)

Colloidal system	Percentage of material in water-stable aggregates larger than 0.1 mm. in diameter	
	Quartz sand	Orthoclase
Ca-Putnam clay.....	28.5	19.0
H-Putnam clay.....	33.5	42.0
H-Putnam 92 per cent + H-straw 8 per cent....	41.0	74.0
H-Putnam 92 per cent + H-straw 8 per cent + Ca(OH) ₂ †.....	45.0	58.0
Ca-Straw.....	71.0
H-Straw.....	94.0

* 1 gm. of colloid and 2 gm. of particles smaller than 0.1 mm. mixed and dried.

† Calcium equivalent to the measured exchange capacity of the system.

of sands and orthoclase into water-stable aggregates. This effect is more pronounced in the presence of hydrogen ions on the exchange complex. It is also evident that the addition of small amounts of organic colloids increases aggregate formation. It is especially significant that Ca-systems are not as highly aggregated as the H-systems. This fact confirms other observations that Ca-humates are more reversible than H-humates (5). It also emphasizes that the beneficial

effects of lime on soil structure cannot be explained solely upon the direct influence of the calcium ion on the flocculation of organic colloids.

The question naturally arises as to the mechanism of the cementation effects of organic matter to produce stable aggregation. At the present moment the exact nature of organic-matter effects is not completely understood. Several explanations have been advanced to explain the tenacity with which granules that are cemented by organic matter hold together.

Williams (50), one of the leaders in the rise of soil science in Russia, has visualized the tenacity of aggregate formation as being due to the saturation of the aggregates with ulmic acid, which is secreted by anaerobic bacteria during the decomposition of the roots of plants. The ulmic acid that soaks into aggregates is considered to pass over into the denatured state as a result of drying or freezing. When divalent cations are associated with ulmic acid, a water-stable cement is supposedly produced. It is significant to note that Williams has not attributed this property to humic, aprocrenic or crenic acid, which he has also considered to be present in soil organic matter.

Sideri (43) has suggested that humus is adsorbed by clay through the process of the orientation of organic molecules on the surface of the clay particles. This adsorption is somewhat irreversible in the moist state; the irreversibility is increased upon dehydration. The data of Myers (35) on the mutual effects of organic and inorganic colloids on the physicochemical properties of a mixture of these colloidal materials point towards polar adsorption as an explanation of the union between the two.

Surfaces possessing electrical properties orient polar compounds. The humic compounds are polar and are therefore capable of being oriented. Since they are only slightly ionized compounds, the carboxyl ends (or the end of the complex with the greatest number of carboxyl groups) are positive. Soil colloids, bentonite, quartz and orthoclase possess electrical properties and, therefore, serve as orienting materials. They are all electronegative and attract toward their surfaces the positive end of a polar compound. The polar adsorption results in a close packing of the organic colloid particles on the surfaces of the orienting substances. It is possible that mineral particles are joined together during this adsorption process. This type of adsorption results in a reduction of the exchange capacity of the system, not because of chemical union but because of steric hindrance to the passage of the ions from the ends and sides of the organic colloids. The polar adsorption not only is a factor in reducing the exchange

capacity of the organic colloid but it also probably impedes the exchange of those cations associated with the inorganic colloids.

It seems, therefore, that dehydration of the adsorbed humus results in a stable union between inorganic and organic materials. Since rehydration is extremely slow, this process causes a tenacious cementation effect which is responsible for aggregate formation.

Kubiena (29) pictures the genesis of the films or binding materials as dependent upon dehydration processes. If the colloidal material in the soil solution is flocculated, evaporation may cause the formation of intergranular braces as the flocculated material accumulates between particles, when the moisture content is gradually reduced. If the colloids are dispersed they may form uniform coatings over the particles as the soil is dried. He visualizes a gradual dissolution of the humus in the slightly alkaline solution that results from the concentration of bases in the final stages of drying. At this stage there is a migration of the concentrated soil solution inside the aggregates towards the surface. The alkali-soluble humic materials then coat the surface and bind the aggregate together when dehydration is complete. This process is considered to be particularly applicable to the formation of chernozem aggregates.

It is obvious that a final explanation of the mechanism of the cementation effects of organic colloids in aggregate formation has not as yet been given. Nevertheless, the majority of the evidence points to some type of oriented adsorption of organic molecules that is stabilized by subsequent dehydration.

IRREVERSIBLE IRON COLLOIDS AND AGGREGATION. Experience in the chemistry of ferric hydroxide has shown that this hydrated colloid becomes almost completely irreversible upon dehydration. There is sufficient evidence to suggest that this irreversibility of colloidal iron hydroxide is the important factor in the production of stable aggregates in certain soils. This is especially true in lateritic soils, that are known both for their high degree of aggregation and for large iron content. Roberts (37) has reported that as high as 71 per cent of the total particles in the lateritic Nipe clay from Puerto Rico consist of aggregates larger than 0.05 mm. in diameter. As much as 95 per cent of the total silt and clay are aggregated into units larger than silt.

Lutz (30) has shown that the lateritic Davidson colloid is flocculated, irrespective of the nature of the cation on the exchange complex. The colloid is only weakly hydrated. Later studies by Lutz (31), on the relation of "free" iron to aggregation, have indicated a very close relationship between the amount of iron and the quantity of water-stable aggregates. He suggests that iron may serve a dual

purpose in aggregation. That part which is in solution may act as a flocculating agent, and the other part which is more gelatinous in nature may exert a cementation action. Most of the iron is undoubtedly precipitated as a hydrated gel at the pH value of the soils that were investigated. Dehydration of these gels should form a rather good cement for binding the flocculated particles together.

It is possible to change the properties of such a highly hydrated colloid as bentonite by treating with iron salts and then dehydrating the system. The bentonite assumes physical properties similar to those of lateritic colloids. The B-horizon of true podsoils is usually aggregated to some extent as a result of the iron and humus compounds that have migrated from the upper layers. Cementation effects of these colloids are responsible for the binding together of the sand and silt grains, generally predominant in these soils.

The importance of colloidal iron in affecting the properties of soils has not been evaluated sufficiently to date. There has been a certain amount of recognition given to its effects upon the base exchange material in soils. It seems plausible to suggest that more emphasis in the future will be placed upon its influence on the physical behavior of soils. It is also possible that colloidal alumina may play a similar role to that of iron.

Relation of Climate to Aggregate Formation. The extent of aggregation in different soils varies considerably. It is of interest to attempt to correlate aggregate formation with the soil-forming climatic factors, rainfall and temperature. The percentage of aggregates in a given weight of soil is affected considerably by the texture. Coarse-textured soils do not have as much silt and clay to aggregate as the finer-textured ones. Consequently, it is important to know not only the total percentage of aggregates in the soil but also the amount of silt and clay, in percentage of the total silt and clay present, which is in the form of aggregates.

Aggregate analyses of a large number of different soils are available which have yielded significant data for a correlation between climate and aggregation (3). Although an insufficient number of samples has been analyzed to formulate accurate mathematical relationships, the results warrant the formulation of several tendencies.

The results are summarized schematically in Figure 33. It is obvious that several striking relationships exist between aggregation, and rainfall, when temperature is kept constant. The percentage of aggregates in a given weight of soil is a maximum in the semi-arid and semi-humid regions, where the chernozem-like and humid-prairie soils are located.

Let us discuss the shape of the percentage of aggregates-rainfall curve. The percentage of aggregates is low in desert soils because of a small clay content. Under arid conditions chemical weathering does not proceed very far; consequently, a small amount of clay is formed from the clay-forming minerals. This small clay content, even though it may be in the form of secondary particles, causes a relatively small number of aggregates to be present. As the rainfall becomes greater, chemical weathering is intensified and, according to the data of Jenny (24), clay formation increases. This increase in the clay content of the surface soil obtains until the rainfall becomes great enough to cause eluviation of the clay from the A- to the B-horizon. The decrease in the clay content of the A-horizon diminishes the possibilities for aggregate formation. Not only is the clay content highest in the chernozem and dark humid-prairie regions but the percentage of organic matter in these soils is also largest. As a result of these two factors, the percentage of aggregates in the soil is greatest in the chernozem and dark humid-prairie soils. For example, the amount of aggregates in podsol, dark humid-prairie, chernozem-like, chestnut-colored and desert soils is approximately 10, 40, 40, 25 and 10 per cent, respectively. The curve in Figure 33 parallels that obtained by Jenny for the clay content of these soils.

If one considers the percentage of silt and clay which is in the form

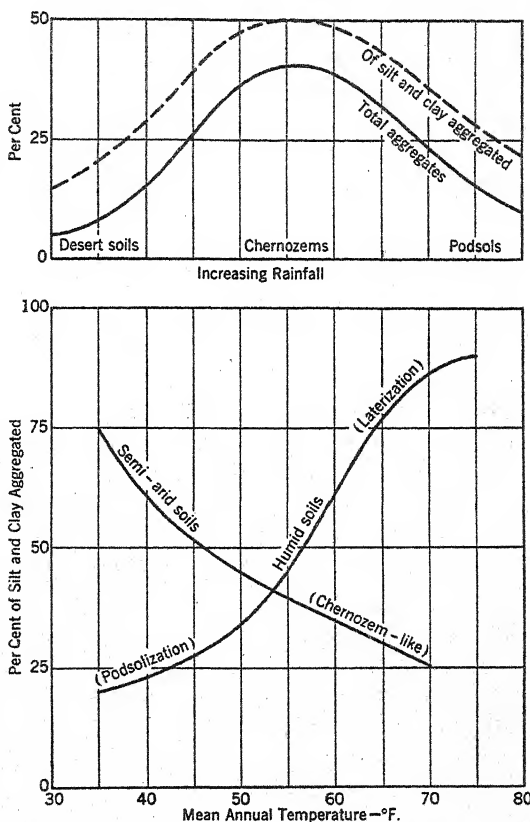


FIG. 33. The climatic aspect of soil aggregation.

of stable aggregates, the effect of texture becomes less important. It is seen that podsol soils have a relatively low percentage of the silt and clay aggregated. Only about 25 per cent of the silt and clay are aggregated. The dark humid-prairie, the chernozem-like soils and the chestnut-colored soils have about 50 per cent of the total silt and clay aggregated. The low extent of aggregation in the podsol type of soils is probably due to the low organic matter content and the removal of alumina and iron from the A-horizons. The small amount of aggregation is the result of some organic matter which is present in the A₁-horizon. That alumina and iron can cause the formation of stable aggregates is evidenced in the extent of aggregation of the coffee-brown layer of the podsol, where about 50 per cent of the silt and clay are aggregated. Of course, organic matter also contributes to aggregation within this layer.

The larger percentage aggregation of the silt and clay in the regions of decreasing rainfall is due to organic matter and divalent bases. The presence of sodium in truly desert soils, together with low amounts of organic matter, causes only a slight aggregation of the small quantities of silt and clay in the soils. Less than 25 per cent of the finer fractions are aggregated.

When rainfall is kept constant and temperature is increased, the percentage aggregation of the silt and clay is affected differently in humid and semi-arid regions. Considering the chernozem-like soils of the semi-arid regions, the percentage aggregation of the silt and clay decreases from Canada to Texas. Since the calcium content is practically constant, this decrease in aggregation is due to a lowering of the organic-matter content with the increasing temperature. This curve is similar to the organic matter-temperature relationship of Jenny (22). The percentage aggregation of these soils varies from 75 per cent in Canada to 25 per cent in Texas.

The humid soils present a somewhat different picture. The percentage aggregation varies from about 25 per cent in the podsols to 95 per cent in the true laterites. The causes associated with the small percentage aggregation of the silt and clay in the podsols have been previously discussed. Aggregation in the laterites is caused by the cementation effects of dehydrated alumina and iron oxides. Thus, in lateritic soils, alumina and iron contribute to aggregation, while in the podsol any secondary-particle formation is due to small amounts of organic matter. Differences in the factors contributing to the building of secondary particles, therefore, are responsible for the S-shaped curve, which correlates the percentage aggregation of the silt and clay in humid soils with temperature.

These tendencies indicate that the soil-forming climatic factors of rainfall and temperature are expressed in the structure of the major soil groups. In addition to these broader effects of climate, there are also the influences of wetting and drying as well as freezing and thawing that contribute to secondary-particle formation.

Processes Contributing to Aggregate Formation. It has been shown that aggregate formation in soils depends upon (1) the presence of small primary particles that may be aggregated, (2) the coagulation or flocculation of the particles and (3) the cementation of the coagulated material into stable aggregates. Even though the nature of the flocculating and cementing agents in soils is fairly well understood, there is no clear picture of the exact mechanism of aggregate formation under natural conditions. Zakharov (53) has presented several of the earlier Russian ideas concerning the genesis of structure. Apparently, the activity of the root systems, together with the soil fauna, was first considered as the primary factor in aggregate formation. This concept was then supplemented by the idea that periodical variations in moisture and temperature caused a fracturation or fragmentation of the soil mass into aggregates.

Later, Gedroiz (14) suggested that aggregate formation is primarily dependent upon pressure and coagulation. This concept has been adopted by Tiulin (46) and others in their work on the genesis of soil structure. Gedroiz visualizes vegetation, and to some extent small burrowing animals, as contributing the major pressure effects. The penetration of roots or animals into the soil causes a pressure adjacent to the channels, and this pressure forces the particles together. Tiulin conceives that pressure produces more intimate contact between particles, so that the cementing influences of the water films are rendered more effective. He recognizes the importance of the root system of plants but attributes a much greater role to the swelling of the soil colloids. The effect of swelling is given particular significance in the development of aggregates in the subsurface horizon. It seems, therefore, that pressure may be important in causing particles to be brought into more intimate contact with each other, but that the exact role of pressure in structure formation is difficult to evaluate.

Gedroiz considers drying, freezing, mutual flocculation by oppositely charged colloids and precipitation by electrolytes to be responsible for the coagulation of the finer mechanical separates in soils. The importance of cations, iron and alumina in aggregation has already been discussed. The effect of drying and freezing will be the subject of subsequent paragraphs. It is essential at this time to point out that the emphasis which the Russian scientists place upon the cal-

cium ion in aggregate formation undoubtedly arises from the experimental techniques they have employed. Most of their conclusions are based upon a comparison of Ca-saturated systems with either NH_4 - or Na-saturated soils. Consequently, calcium does have a marked direct effect upon the granulation of non-acid soils.

ALTERNATE WETTING AND DRYING. Common experience has taught us that compact, tightly cemented clods will slake down into smaller aggregates as a result of alternate wetting and drying. The farmer likes to have his "clay land" dry out thoroughly and then be re-wetted slowly in order to produce a seedbed that has good tilth. What are the causes of this aggregate-forming effect of alternating wetting and drying?

Most of the experimental evidence shows that drying or dehydration of the soil colloids causes a shrinkage of the soil mass and a cementation of clay particles. It should be obvious that the dehydration of the soil cannot be uniform, especially if the drying process is rapid. Consequently, throughout the mass, unequal strains will arise which tend to form clods. This difficulty of even drying, without the production of cracks, is almost always encountered when one attempts to desiccate a puddled ball of clay.

At least two processes operate to cause disruption of the clod into smaller units when the dried clod is wetted. The rapid intake of water causes unequal swelling throughout the clod, which produces fracturation and fragmentation along the cleavage planes. Moreover, the sorption of water into the capillaries results, first in a compression of the occluded air, and finally in a virtual explosion within the clod, as the pressure of the entrapped air exceeds the cohesion of the particles. Yoder (52) has been one of the first to recognize the importance of entrapped air in the disruption of large aggregates and clods during the wetting process. His hypothesis is supported by the facts that wetting in a vacuum or slow wetting by capillarity does not cause violent disruption of the clod. When the clod is wetted slowly, the air is expelled from the larger pores as the smaller ones take up water by capillarity. If the clod is immersed in water, disintegration into smaller fragments takes place as the air is expelled. This disintegration of the clod is almost concurrent with the rise of air bubbles to the surface of the water.

Henin (19) has attached much importance to the Yoder concept and has attempted to develop a mathematical expression for defining the stability and instability of aggregates and clods in water, in terms of capillary and cohesive forces. The condition for disruption is given by the expression; $rC + C_1 < 2A$, where r is the diameter of the

largest capillaries in the soil, C is the apparent cohesion of the soil, C_1 is the cohesion of water and A is the affinity of the soil for water. Disruption is considered to take place only with the larger aggregates that possess capillaries in which the air can be compressed. Henin considers both swelling and disruption as factors for the disintegration of clays on wetting. It is significant to note that disruption is decreased by the addition of alcohol to water. The amount of aggregates larger than 0.2 mm. in diameter is correlated with the affinity of the soil for the liquid rather than with the surface tension of the different alcohol-water mixtures. This fact emphasizes the significance of unequal hydration and swelling in the slaking of desiccated clods.

It is a common experience that an aggregate analysis of air-dried samples of soils, containing much clay, apparently gives too many aggregates. These same soils show little evidence of granulation under natural conditions in the field. If these air-dried soils are placed in alcohol, there is little slaking and few smaller particles are formed. It seems, therefore, that the fragments resulting from the disruption of large, compact lumps or clods of clay are not true granules, but represent only a particular condition of aggregation that depends upon the state of dehydration of the soil. In most instances, this so-called granular condition reverts to a more or less non-aggregated state, after sufficient moisture is present to rehydrate the particles thoroughly and permit maximum swelling to take place throughout the soil mass.

Haines (16) has developed an interesting picture of the variations in soil structure that occur as a result of volume changes associated with varying moisture content. His investigations shed considerable light on the importance of shrinkage and swelling, which result from alternate drying and wetting, in the formation of good soil tilth. His method consists in measuring the volume of a puddled block of soil as it is dried and rewetted. When the volume of soil is plotted as a function of the volume of water removed, several very significant facts concerning shrinkage become obvious. A few important curves from his original data are shown in Figure 34. Curve A represents a clay separate containing 90.5 per cent clay, curve B is for kaolin containing 52.8 per cent clay and curve D represents a clay subsoil that has been alternately dried and rewetted.

It is noted that, as the thoroughly puddled samples are dried, the decrease in the volume of the soil is equal to the volume of the water lost. All the curves are parallel in the wet region and have a slope of 1. As dehydration progresses, there is a distinct break in the curve and the change in soil volume becomes much less than the volume of

water removed. This break signifies the point at which air enters the soil. The shrinkage over the lower portion of the curve varies with the nature of the soil. Haines has suggested that this portion of the

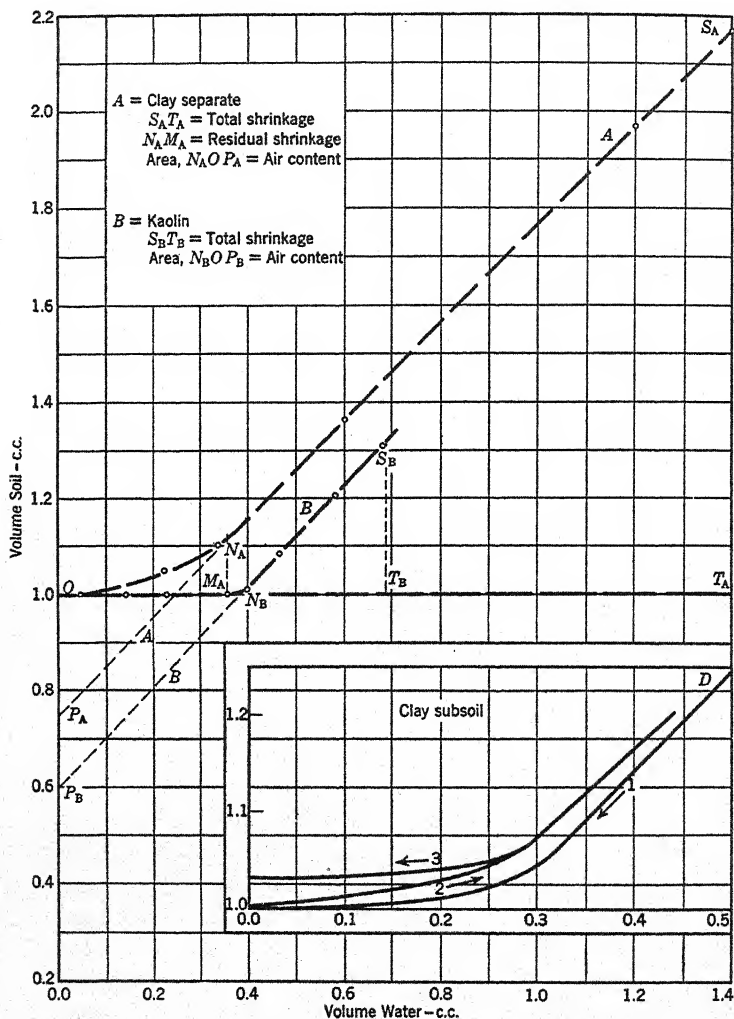


FIG. 34. The shrinkage of soils as a function of moisture (Haines, 16).

curve be called "residual shrinkage," as distinguished from the total shrinkage that takes place from complete saturation to dryness. If the straight portion of the curve, where the volume change is proportional to the water loss, is extrapolated to intersect the y-axis, then the

values of OP_A and OP_B will represent the volume of the pore space that is occupied by air in the dried soils.

The total porosity of A and B is 22.5 and 37.8 per cent, respectively. The clay separate has a total shrinkage of 130 per cent (S_AT_A) and a residual shrinkage of 8.6 per cent (N_AM_A). Kaolin has a total shrinkage of 30 per cent (S_BT_B) and no residual shrinkage. Total shrinkage appears to be dependent upon the clay contents. Residual shrinkage varies with the hydration of the soil colloidal material, both organic and inorganic.

It is important to compare the alternate drying and wetting curves of subsoil D . It is observed that the volume of the rewetted soil is greater than that of the original. This increase in volume is permanent after complete dehydration and is undoubtedly due to the entrance of air into the soil mass; the air then becomes occluded in the pore spaces. These particular curves show that the drying process must extend below the point at which air enters the soil before rewetting occurs in order to obtain an increase in porosity. They also lend emphasis to the significance of the compression and release of the entrapped air as a possible method of clod disintegration.

Summarizing, it is apparent that alternate wetting and drying produce aggregation as a result of unequal strains and stresses that are set up by shrinkage and swelling processes, together with the disruptive action of air entrapped in the pores on wetting. Drying causes a cementation of the clay particles as the soil mass shrinks. Air enters into the pores during dehydration. The rapid intake of water during wetting is responsible for unequal swelling as well as the compression of the occluded air. These two effects cause a crumbling of the clod along numerous cleavage planes. Repetition of this process accomplishes a high degree of disintegration of the cohesive lumps and clods. This secondary-particle formation, however, is not exactly permanent, since an excess of water may cause rehydration of the entire soil and a return to a non-aggregated condition. It is highly probable that an aggregate analysis of a dried clay soil often gives a misleading picture of the natural aggregation.

ALTERNATE FREEZING AND THAWING. The farmer recognizes the beneficial effects of freezing and thawing upon soil tilth when he plows clay soils in the fall or late winter. Fall-plowed clays are friable and easily tilled in the spring. Alternate freezing and thawing cause a granulating action on soil clods that is usually more effective than drying and wetting. It must be stated, however, that freezing and thawing do not always result in a highly aggregated soil in the spring. Certain moisture conditions seem to be essential for the maximum

effects of freezing. If the soil is dry during the winter, there will be little disintegration of the clods. If the soil is excessively wet and thawing is accompanied by rain, there may be dispersion of any aggregated material. It is often observed that the aggregation produced by several freezings and thawings decreases rapidly as the number of freezings increases. This is particularly true of silty soils that are low in organic matter. Similar to the aggregates produced by alternate wetting and drying, those that are formed by freezing and thawing are more or less temporary unless sufficient organic matter is present to stabilize them. That is, the major beneficial effects of freezing and thawing are most evident during the early part of the spring season and on soils that contain appreciable amounts of organic matter.

Jung (25) has made an extensive study of the nature of frost action on soils and has shown that freezing may cause either aggregation or dispersion. The nature of the crystallization of ice is the determining factor. Crystallization is influenced by the rapidity of cooling. With slow cooling, the ice crystals form in the tension-free pore spaces. These crystals serve as growth centers to which water is drawn from around the particles. This process causes a dehydration of the particles, which makes possible more intimate contact between the soil grains. Moreover, there is a possibility of a certain amount of flocculation, due to any electrolytes that might be released during dehydration.

The aggregation by mechanical means that results from the pressure of the ice crystals is probably more important than these effects. Slow cooling brings about a relatively small nucleus formation and large crystals. These crystals partially melt during thawing and then serve as nuclei for further freezing. The pores that are enlarged by these crystals cause a loosening effect upon the soil. In addition, the combined influences of crystal pressure and dehydration produce pronounced aggregating effects.

If the cooling is rapid, large numbers of small crystals are formed which cause a breaking up of aggregates, by the expansion of water. The same principle of quick freezing applies here as in the manufacture of fine-textured ice cream.

The highly interesting results of Jung are illustrated in Figure 35. It is noted that quick freezing at -190°C . has the opposite effect of slow freezing (-10°C .) up to about complete moisture saturation, where the two curves meet. The quickly frozen soil has about the same degree of dispersity as the unfrozen at moisture contents below about 40 per cent saturation. It is apparent that the water content at freezing is as important as the rate of cooling. Up to about 50 per cent

saturation, when the soil is in the gel state, aggregation under slow cooling increases with moisture content. Here, a small number of large crystals is formed as compared with a large number of small crystals under rapid cooling. The soil becomes more fluid above 50 per cent saturation, and aggregation under slow freezing decreases until it reaches the same value as that of rapid freezing. These data point out the explanation of the decrease in aggregation that occurs under natural conditions when freezing and thawing take place in saturated soils.

ROOT EFFECTS IN AGGREGATION. Two facts stand out as significant concerning the effect of the root system of vegetation on the genesis of soil structure. First, there is a growing belief that plant roots are as important as any other factor in the production of stable granulation. This belief is supported by numerous experimental data and field observations. Second, no satisfactory explanation exists that gives a complete picture of the nature of these beneficial root effects.

Practical experiences and scientific investigations have clearly demonstrated that sod crops promote granulation. Every student of soils is familiar with the high state of granulation of prairie soils. Anyone who has spaded or worked with grass sods undoubtedly has been impressed by the granular nature of the soil among the roots. Evidence of the root effect of a bluegrass sod on granulation is shown in Plate 2. This picture represents a section of bluegrass sod at a depth of about 6 inches, with the exposure parallel with the surface. The entire soil mass is penetrated by countless roots. The rounded granules are literally enmeshed by the root system. This condition approaches the ideal as far as stable granulation is concerned.

What mechanism has been responsible for this type of granulation? There are several explanations that may be offered, although the exact cause or causes cannot be definitely established. As has been previously

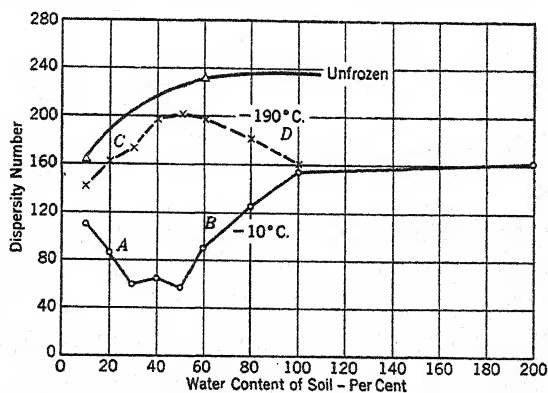


FIG. 35. The dispersion and aggregation of soil as a result of slow and quick freezing (Jung, 25). A—soil in gel state, small number of large crystals; B—soil in sol state; C—soil in gel state, large number of small crystals; and D—water withdrawn from particles.

mentioned, the earliest explanation was based on the pressure exerted by the growing roots, which effects a separation of the particles adjacent

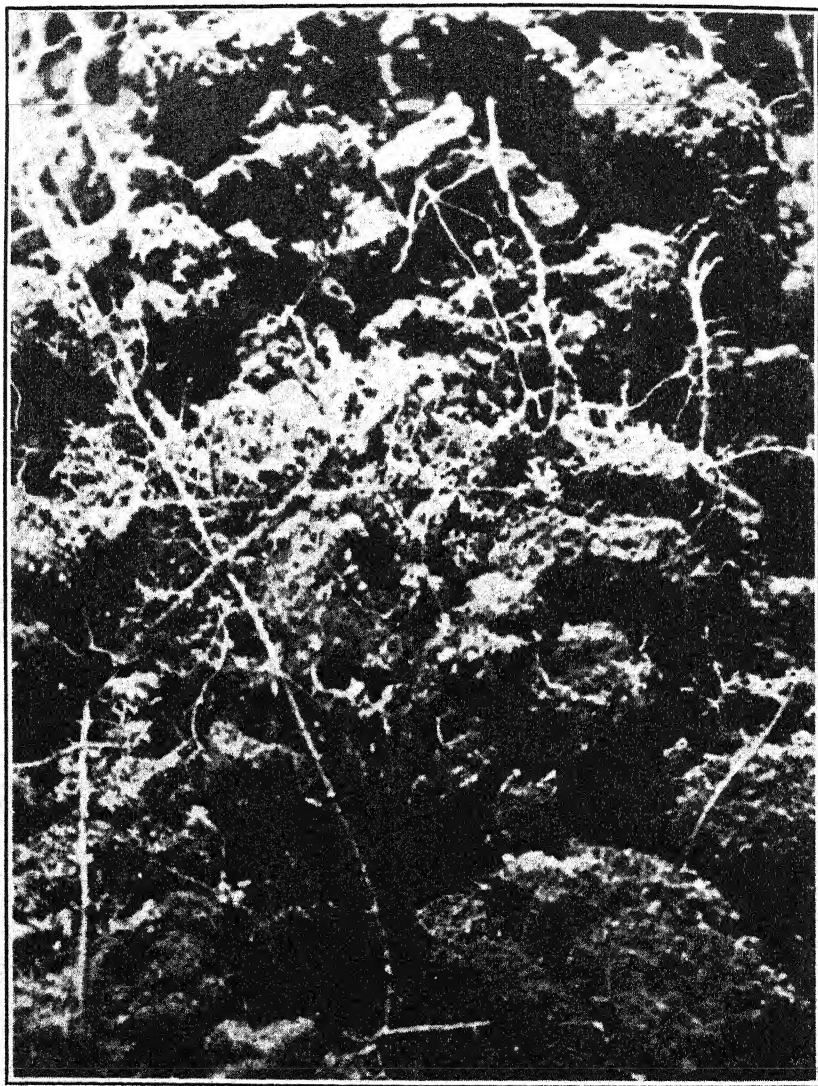


PLATE 2. Photomicrograph of the effect of bluegrass roots on soil granulation (magnification about 10 times).

to the root and a pressing together of these units into aggregates. In other words, each root hair that penetrates into a clod introduces a point of weakness in the clod. The penetration of sufficient root hairs

or roots throughout the clod causes the formation of granules. One of the more recent suggestions is that granulation is accomplished by changes in moisture in the vicinity of the root system as a result of water intake by the plant. This produces localized dehydration, which brings about shrinkage and the formation of surfaces of fracture. It is easily conceivable that both pressure and localized dehydration may be important factors in aggregate formation. However, it is highly probable that other influences are equally as important.

The question of root secretions should not be overlooked, although little experimental evidence exists upon which to formulate definite suggestions. Williams has attributed the high state of granulation of grassland soils to the ulmic acid which is secreted by anaerobic bacteria during the decay of plant roots. However, the presence of this acid in soils has not been universally accepted by soil scientists. There is no reason to doubt the possibility of root excretions that may have definite flocculating or cementing effects on soil particles. This is one field that needs rather thorough investigation.

It is interesting to note that the soil particles within the volume occupied by the roots of the rice plant seem to be aggregated, even though the land is flooded. The clay soil between the plants does not exhibit this aggregation. It would appear that the localized dehydrating effects resulting from water intake could not be important under water. Was the aggregation effected simply by the pressure caused by the penetration of roots through the soil or were root secretions responsible? This question must remain unanswered until further experimental results are available.

It seems plausible to conclude, however, that something comes from the plant root that stabilizes any aggregation that is formed, irrespective of the mechanism of this formation. It may be the organic colloids that are produced during root decomposition. Since it has been shown that organic matter is a stabilizer of soil granulation, it is obvious that the organic residues from plant roots may have considerable effect upon making stable secondary particles, whether the original unit is formed as a result of pressure, dehydration or coagulation. Geltzer (15) explains the larger amount of stable aggregates resulting from sod crops on the basis of the rate of decomposition of the organic matter. Russell (40) interprets this suggestion as meaning that some by-product, probably mucus, is formed during decay. The more rapid the decomposition of cellulose compounds, the greater will be the formation of these sticky substances. This is only a hypothesis but it does give recognition to the possibility of the production of certain stabilizing materials during root growth and decay.

FACTORS TO BE CONSIDERED IN SOIL-STRUCTURE RELATIONSHIPS

In studying soil-structure relationships, one is primarily interested in the degree to which soils exhibit a given structure, the stability of this structure and the physical properties of the soil that are dependent upon the particular structural arrangement.

The Degree of Manifestation of Structure. Soil structure has been defined as the arrangement of particles into certain patterns. The type of arrangement varies with the amount and nature of the secondary particles or aggregates. Between the particles in any soil is the pore space. These pores may be large or small, continuous or discontinuous, depending upon the type and arrangement of the particles. It is obvious, therefore, that differences in soil structure will be expressed by (1) the structural pattern of the various horizons of the profile, (2) the extent of aggregation and (3) the amount and nature of the pore space. Using and expanding the concept of Zakharov, these properties may be used to denote the degree of manifestation of soil structure by a given soil.

Soils vary considerably in the degree to which structure is manifested. The different horizons in the same profile generally do not exhibit the same type of structure or structure capacity. In the discussion of the classification of soil structure it was shown that the structural pattern is primarily dependent upon the type of cleavage and the nature of the aggregates. Thus, the first visual concept of the degree of manifestation of structure in a given profile is the extent and type of cleavage in the various horizons and the nature of the aggregates resulting from such cleavage.

Certain soils exhibit no cleavage and no secondary units; these are the so-called structureless soils. They possess no capacity to break up into aggregates. Other soils show distinct cleavage. It may be directional, horizontal or vertical, or non-directional with a haphazard occurrence of the cleavage planes. When the majority of the cleavage planes are horizontal, the soil is considered to manifest a plate-like structure; if the planes are vertical the structure is columnar or prismatic; and if the planes of fracture run in all directions the structure is usually cube-like in nature. These manifestations, which are related to size, shape and arrangement of particles, are visible to the eye and can be expressed rather easily.

Extent of Aggregation of Soils. The number of secondary particles is a measure of the capacity of the soil to break up into fragments or

its state of aggregation. Although a relative idea of the state of aggregation may be obtained by observing the extent to which the soil crumbles on handling, a much better index is provided by an aggregate analysis of the soil. Such an analysis measures the relative distribution of the various sizes of aggregates and permits a calculation of the percentage aggregation of the finer mechanical separates. These two values are essential to furnish a complete picture of the structure capacity of the soil. It must be emphasized, however, that any measure of the extent of aggregation gives only the number of aggregates and not their arrangement. Consequently, an aggregate analysis is really an index of structure and does not characterize the type of structure.

The term "state of aggregation" has been suggested to designate the percentage of aggregates in a given weight of soil. It is obvious that sandy soils cannot contain as many aggregates as a well-granulated silt loam, because they possess such a high content of coarse mechanical separates. Nevertheless, it is possible for all the silt and clay in coarse-textured soils to be present in the form of aggregates. Therefore, in order to have a measure of the percentage aggregation of the fine mechanical separates, a value is used which is obtained by dividing the percentage of aggregates larger than a given size (0.05 to 0.1 mm.) in the soil by the percentage of mechanical elements smaller than this size. This gives the "degree of aggregation" of the fine particles. In other words, it represents the percentage of particles smaller than a given size that are aggregated into stable units larger than this size.

Examples of the structure capacities of several widely different soils are given in Table 18. These data point out that chernozems and lateritic soils possess a high state of aggregation. Likewise, a high percentage of the small particles is aggregated. Podsoles, however, are poorly aggregated. This is especially true in the A_2 -horizon. The effects of organic matter and iron are noticeable in the A_1 - and B-horizons, respectively. The importance of recognizing the extent to which the smaller mechanical separates are aggregated is shown by comparing the surface layers of the two chernozem-like soils. One is a fine sandy loam and the other a silt loam. The state of aggregation of both is about the same in spite of the differences in texture. This is due to the fact that 75.5 per cent of the particles smaller than 0.05 mm. in the fine sandy loam are aggregated, whereas only 50 per cent are aggregated in the silt loam. It so happens that the photomicrograph *A* in Plate 1 is a structural unit from the surface of this fine sandy loam profile. The picture illustrates why a high degree of aggregation is found by an aggregate analysis.

TABLE 18

THE STRUCTURE CAPACITY OR EXTENT OF AGGREGATION OF VARIOUS SOIL PROFILES

Soil type	Depth	Mechanical separates larger than 0.05 mm.	State of aggregation Aggregates larger than 0.05 mm.	Degree of aggregation Mechanical separates smaller than 0.05 mm. aggregated into units larger than 0.05 mm.
	inches	per cent	per cent	per cent
Chernozem (fine sandy loam)	0-3	41.1	44.2	75.5
	3-9	55.3	30.4	68.0
	9-15	55.2	32.5	72.1
	18-28 (lime)	37.1	33.6	53.4
Chernozem-like silt loam	0-6 (A ₁)	13.3	43.4	50.0
	7-12 (A ₂)	7.6	70.9	76.5
Lateritic clay	0-5	29.1	67.1	94.6
	5-30	14.1	70.6	82.1
	30-60	28.7	37.7	52.9
Podsol (fine sand)	0-4 (A ₁)	72.8	8.3	24.5
	4-10 (A ₂)	64.8	3.9	11.1
	10-16 (B ₁)	77.5	10.9	51.2
	16-28 (B ₂)	85.7	7.7	53.6
	28-36 (B ₃)	85.6	8.1	35.8
	36+ (C)	74.5	0.7	26.8

Soil Porosity. The arrangement of the soil particles determines the amount and nature of the soil pores. Soil porosity may be defined as that percentage of the soil volume which is not occupied by solid particles. In a soil containing no moisture, the total pore space will be filled with air. The pores of a moist soil are filled with both air and water. The relative amounts of air and water present will depend largely upon the size of the pores. The size of the pores is dependent upon the size and character of the arrangement; pore size, however, cannot foreshadow a granular, nutty or columnar structure of the soil.

The porosity of the soil is calculated from the real and apparent specific gravity. The real specific gravity of the solid particle is determined by means of a pycnometer or specific-gravity bottle. It represents the weight of one cubic centimeter of solid particles. The apparent specific gravity or volume weight is measured by weighing

a given volume of soil in its natural structure. It represents the weight of one cubic centimeter of soil and pore space. Naturally, since only the solid particles in a dried soil contribute to this weight, it is possible to make an easy calculation of the total pore space. This is done by the well-known formula:

$$\text{Per cent pore space} = \left(1 - \frac{\text{apparent specific gravity}}{\text{real specific gravity}}\right) \times 100$$

This expression gives the volume percentage of the pore space, but does not characterize the size of the pores. The total porosity of an average soil varies in the neighborhood of 50 per cent. Sands usually have less than this amount; clays and organic soils have higher porosities. The wide variations in total porosity are shown by the data in Table 19. It should be obvious that porosity will vary with the size of the particles and the extent of aggregation.

TABLE 19
THE TOTAL POROSITY OF VARIOUS SOILS

Type of soil	Depth inches	Total porosity, per cent by volume	Investigator
Chernozem-like fine sandy loam.....	0-3	60.6	Baver, <i>et al.</i> (<i>Amer. Soil Survey Assn. Bul.</i> 14, 1933)
Hardeman clay loam.....	1½-10	44.4	
Poncena clay.....	0-7	49.8	
Nipe clay.....	0-5	56.6	
Georgeville silt loam.....	½-8	57.0	
Kirvin fine sandy loam.....	at 6	35.7	Middleton, <i>et al.</i> (<i>U. S. Dept. Agr. Tech. Bul.</i> 430, August, 1934)
Nacogdoches fine sandy loam.....	at 5	43.4	
Vernon fine sandy loam.....	at 2½	41.5	
Muskingum silt loam.....	at 3½	45.1	
Muskingum silt loam.....	at 3½	31.3	
Cecil sandy clay loam.....	at 3	45.3	
Shelby silt loam.....	at 3½	45.4	
Shelby silt loam.....	at 13	51.7	
Shelby silt loam.....	at 54	31.7	
Palouse silt loam.....	at 10	50.2	
Colby silty clay loam.....	at 5	48.9	
Clinton silt loam.....	at 4½	51.9	
Marshall silt loam.....	at 6½	56.4	
Marshall silt loam.....	at 18½	60.3	
Houston black clay.....	at 4	61.9	
Houston black clay.....	at 35	40.2	

The total porosity is not as important for characterizing the structural properties of soils as the relative distribution of the pore sizes. Clays, for example, tend to have higher total porosities than sands. Everyone is aware, however, that the porosity relations of these two types of soils are distinctly different. Clays possess a large number of small pores which contribute to a high water-holding capacity and slow permeability. Sands have a small number of large pores which are responsible for rapid drainage and a low moisture-holding capacity. Schumacher (see Chapter I) suggested that the relative proportion of capillary to non-capillary pores is responsible for optimum soil-structural properties. The non-capillary porosity is the sum of the volumes of the large pores, which will not hold water tightly by capillarity. They are normally filled with air and are responsible for the air capacity and ready percolation of water through the soil. Capillary porosity is the sum of the volumes of the small pores that hold water by capillarity. They are responsible for the water capacity of soils.

Doiarenko (28) has developed this concept of Schumacher's rather extensively. He and his coworkers have investigated the relationship of non-capillary porosity to various soil properties. Some of their important results are given in Table 20. These data were obtained

TABLE 20

THE RELATIONSHIP OF SOIL POROSITY TO THE SIZE OF AGGREGATES (DOIARENKO, 28)

Soil property	Diameter of aggregates in mm.				
	Smaller than 0.5	0.5-1.0	1.0-2.0	2.0-3.0	3.0-5.0
Total porosity, per cent.	47.5	50.0	54.7	59.6	62.6
Non-capillary porosity, per cent. ...	2.7	24.5	29.6	35.1	38.7
Capillary porosity, per cent.	44.8	25.5	25.1	24.5	23.9
O ₂ content of soil air, per cent.	5.4	18.6	19.3	19.4
O ₂ content of soil, per cent.	0.1	4.5	5.7	6.7	7.5
Nitrate formation, mgm. N per kg. soil.	9.0	19.1	34.0	45.8

with aggregates of different sizes that were removed from the soil by sieving. It is noted that as the size of the soil granules decreases from 5 to 0.5 mm. the total porosity falls about 24 per cent. The really significant change, however, is the 93 per cent lowering of the

non-capillary porosity. The air capacity decreases and the water capacity increases. It is important to point out that the oxygen content of the soil air as well as nitrate formation are closely related to the content of large pores. Other recent experimental results have shown rather definite relationships between non-capillary porosity and soil permeability as well as the root development of crops.

The ideal soil should have the pore space about equally divided between large and small pores. Such a soil would have sufficient aeration, permeability and water-holding properties. A clearer picture of soil porosity is given by the two profiles in Figure 36. There are several significant points in this figure that should be emphasized. In the first place, it is seen that the Marshall silt loam has a uniform

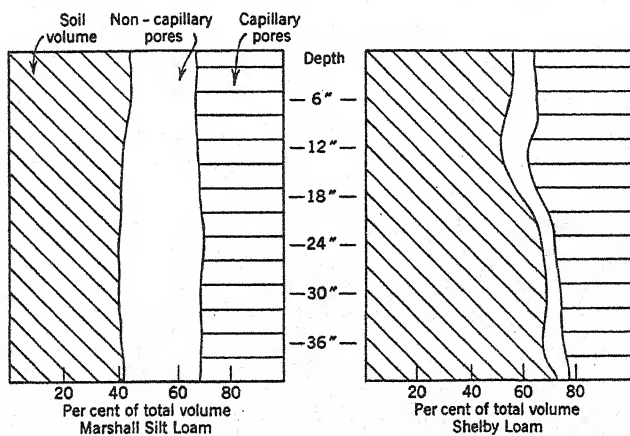


FIG. 36. Pore-space relationships in the Marshall silt loam and Shelby loam.

capillary and non-capillary porosity throughout the entire profile. This is not true of the Shelby loam since the maximum porosity occurs in the upper 12 inches. The Shelby soil has a higher soil volume than the Marshall, which is indicative of its greater apparent density. The most impermeable layer in the Shelby occurs at about 22 to 26 inches, where the non-capillary porosity is only about 5 per cent of the total volume. The minimum non-capillary porosity of the Marshall is about 25 per cent. This makes the ratio of the non-capillary porosity of the Marshall to that of the Shelby equal to about 5.

Approximately 50 per cent of the total pores in the Marshall silt loam are non-capillary. The Shelby, on the other hand, has only a total porosity of about 25 per cent in the subsoil, 89 per cent of which are capillary pores. Its water capacity is too high for its air capacity to insure adequate aeration and drainage.

One of the major criticisms of the concept of non-capillary porosity is the fact that the tension of the capillary column at which capillary saturation is achieved is usually not considered. In most cases, however, saturation includes all those pores that will lift water at least 10 cm. Recent studies on the relation of capillary tension to non-capillary porosity will be discussed in Chapter VI. The expression of the soil-pore space in terms of air and water capacity is very simple and affords a means of interpreting a rather complicated physical property in such a way that its practical implications are easily visualized.

The Stability of Structure. The stability of structure refers to the resistance that the soil aggregates offer to the disintegrating influences of water and mechanical manipulation. The stability of the aggregate is of utmost importance in forming and preserving good structural relationships in soils. Rather compact and coherent aggregates may be found in the dry state, but, if these secondary particles disintegrate in water, the aggregation is not very stable. Water may cause the deterioration of aggregation in two ways. First, there is the hydration effect of the water which causes a disruption of the aggregate through the processes of swelling and the exploding of entrapped air. Henin (19) considers these two factors as dominant in the degradation of soil structure. The air-dried aggregates are not stable in an excess of water but swell and then disintegrate to give a compact arrangement of particles. It is a common occurrence for clay soils, especially heavy subsurface layers, to become aggregated to a certain extent under conditions of extreme drought followed by gentle rains. The soil is friable and permeable. However, if a prolonged wet period follows and the aggregates are subjected to excessive wetness, complete hydration of the aggregate ensues. Swelling produces a compactness of the soil particles and the permeability of the soil is markedly decreased. This type of disintegration occurs with those aggregates that are formed by alternate wetting and drying or freezing and thawing. Such aggregates are generally unstable under conditions of maximum hydration. On the other hand, aggregates that are cemented or stabilized with organic matter or colloidal iron compounds are very resistant to decomposition by the hydration effects of excess water.

The second manner in which water destroys aggregation and deteriorates soil structure is by falling rain. The impact of falling drops of water on exposed soil exerts a significant dispersing action on the aggregates. The dispersed particles are then carried into the soil pores, causing increased compaction and decreased porosity. This deteriorating effect upon soil granulation has been recognized by most

soil physicists since the time of Schübler. Wollny (51), in 1874, observed that rye, peas and vetch protected the soil from raindrops to such an extent that the content of large pores was 34 to 53 per cent higher than on an adjacent unprotected soil. This clogging influence of dispersed particles on soil porosity has been recently confirmed by several investigators. Intense applications of rain destroy the granulation and open structure of the top inch or more of the soil, so that there is a dense and impervious surface. This type of structure degradation is least common with those aggregates that are stabilized with humus or iron compounds. It is important to remark that sufficient emphasis has not been placed upon the deteriorating action of falling raindrops. In many instances raindrops are the major cause of the dispersion of soil aggregates. Their immediate influence is confined to a shallow layer in the surface, but the structure of this layer may be so broken down as to limit the air and moisture relations of the entire profile.

Cultivation and other tillage operations generally tend to promote a continued decrease in soil granulation. The fact that land is cultivated means that the soil is more exposed to the action of raindrops, to increased organic matter decomposition and to the removal of indirect structure-forming materials. Thus, cultivated soils are less granular than the corresponding virgin areas. Moreover, mechanical manipulation of the soil, especially at the wrong moisture content, almost always leads to a crushing of the soil aggregates. Although the significance of tillage on soil properties will be discussed later, it should be mentioned in passing that the cultivation of soils may not lead to an extensive degradation of structure if the organic-matter content of the soil is maintained at a relatively high level.

The stability of soil aggregates is readily determined by measuring the decrease in aggregation as a function of the time of shaking in water. For example, the degree of aggregation of the Grundy and Marion silt loams was 51 and 22 per cent, respectively, if the samples were analyzed without shaking. Shaking for ten minutes in a reciprocal shaker reduced these amounts to 41 and 11 per cent, respectively; the reductions corresponded to about 20 and 50 per cent, respectively, of the silt and clay that were aggregated. One hour of shaking caused a 53 and 86 per cent, respectively, decrease in the degree of aggregation. These data point out that not only is the Grundy silt loam more highly aggregated than the Marion silt loam but that its granules are also more stable.

Soil Properties Dependent upon Structure. As previously stated, it is essential to distinguish between soil structure (the arrangement

of particles) and soil properties that depend upon structure. Porosity, for example, in the true sense, is a property of the soil that is chiefly dependent upon particle arrangement, even though the amount and nature of the pores have been considered under the category of the degree to which soils exhibit structure. The nature of the soil is intimately associated with the water and air relationships. The capacity of the soil to hold water and air as well as the movement of air and water through soils are important properties that are primarily functions of structure. Either of these factors may serve as an index of structure. Water-structure relationships will be discussed in Chapter VI and water-air relations in Chapter VII.

METHODS OF EVALUATING SOIL STRUCTURE

There are direct and indirect methods for evaluating soil structure. The direct methods, macroscopic and microscopic observation, aim to portray the character of particle arrangement, the size of aggregates and the type of cementation. Indirect methods are usually based either on a measure of the extent and stability of aggregation, the amount and distribution of the soil pores or the permeability of the soil for water.

Macroscopic and Microscopic Methods. No macroscopic or microscopic method has been sufficiently perfected to date to provide a complete picture of soil structure. Progress has been made in attempting to photograph (1) the non-capillary pore space that can be filled by various fixing liquids and (2) the appearance of the mineral grains in relation to cementation agents. Nevertheless, no satisfactory technique is available to picture the arrangement of the soil particles as they exist in the various types of structure.

Pigulevsky (28, 36) has devised a method of soil fixation whereby thin sections can be examined under the microscope. A mixture of three parts of paraffin and one of naphthalene is forced into the soil-pore space. After cooling, the mass can be cut and polished for study under the microscope. Harper and Volk (17) have modified Pigulevsky's technique by substituting lacquer or Bakelite varnish as the cementing material to fill the pores. Thin sections are prepared and photographs are taken to give a magnification of 4.5 diameters. The non-capillary porosities of two distinctly different soils are shown in Plate 3. It is seen that the Marshall silt loam profile exhibits a rather uniform non-capillary porosity throughout all horizons. This is in accordance with the porosity data in Figure 36. The Parsons very fine sandy loam has a high percentage of non-capil-

MARSHALL SILT LOAM

PARSONS VERY FINE SANDY LOAM

Depth

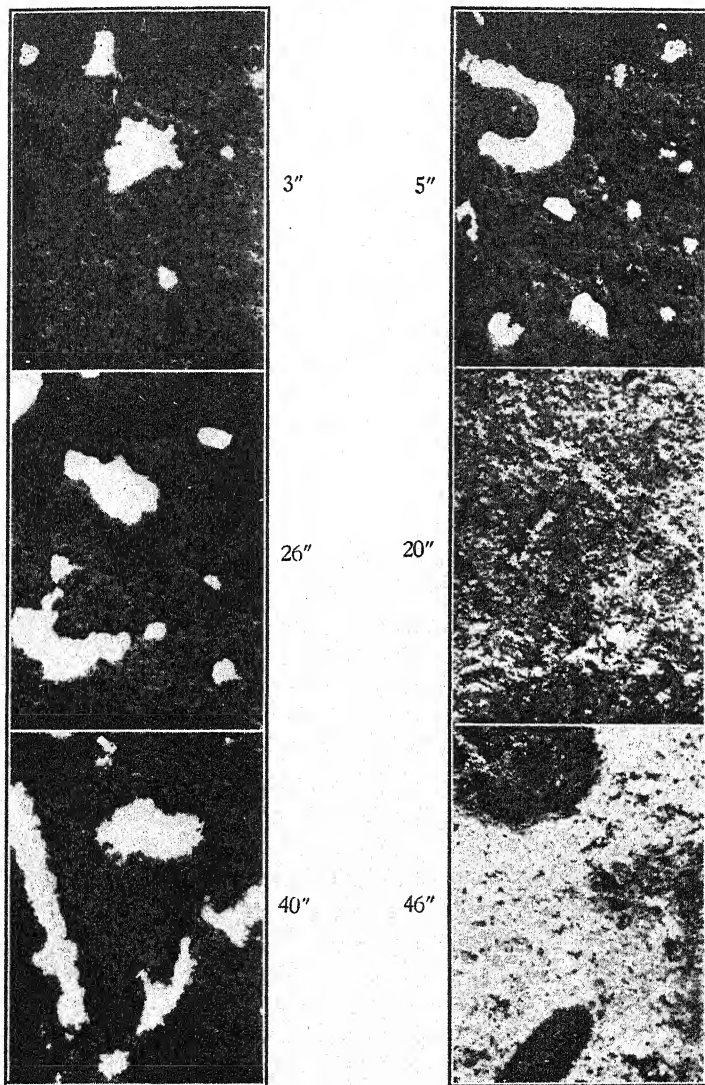


PLATE 3. Photomicrographs of non-capillary pore space in soil profiles (magnification 22 times). (Courtesy of H. J. Harper.)

lary porosity in the surface layer, but very little in the subsoil horizons. These layers give an appearance similar to that in Plate 1, *B*. The particles are so tightly packed that little granulation and non-capillary porosity are perceptible. These photomicrographs show the many possibilities of this type of microscopic study for characterizing the nature of the large pores in the soil. Other physical factors such as the distribution of concretions (see 46 inches depth in Parsons profile), organic matter, etc., can be studied by means of this technique.

Kubiena (29) has perfected special microtechniques for making microscopic studies of soil. A special microscope, using incident light, is employed for making photographs of the soil, either in the field or laboratory. In the field, the microscope is usually mounted against the vertical face of an exposed profile. In the laboratory, thin sections and debris preparations are used. Thin sections provide information of the interior construction of a soil, such as the pore space between particles, root or worm channels, density of packing, etc. Debris preparations consist of a crushed fragment of dry soil that is embedded in Canada balsam. Such preparations show the "arrangement of the soil particles on the surface of the mineral grains" and the "structure of the coatings or the localized grain deposits." Kubiena has suggested that various types of structure can be classified on the basis of the arrangement and affinity of the cementation agents ("fabric plasma") to the mineral particles ("fabric skeleton"). The reader is referred to Kubiena's book for full details of this type of microscopic investigations.

Even though the investigations of Pigulevsky, Kubiena and Harper have provided excellent visual portrayal of certain phases of soil structure, some method of representing the various types of structural arrangements as they appear to the eye is badly needed. It should be possible to photograph the structure of the different soil horizons as they exist under natural conditions and to magnify the detail so as to make the resulting picture show more than the eye can normally see. Such magnification should not be large. In addition to obtaining a visual reproduction of the nature of the cleavage planes and particle arrangement in a given horizon, the type of fragmentation that occurs when the soil breaks up should also be studied. The photographs in Plate 1 are examples of some of the possibilities of this type of approach to the structure problem.

Aggregate Analyses. **TECHNIQUES EMPLOYED.** An aggregate analysis aims to measure the percentage of water-stable secondary particles in the soil and the extent to which the finer mechanical separates are aggregated into coarser fractions. In general, three techniques

are employed to accomplish such an analysis. They are wet and dry sieving, elutriation and sedimentation.

Direct dry sieving of soils as they occur in the field has been used by Keen (27), Cole (11) and others to evaluate the distribution of clods and aggregates. Cole believes that air-dry sifting gives a better picture of aggregation of arid California soils than wet sieving, since the aggregates are so weakly held together in the moist condition that the mechanical action of sieving is sufficient to destroy them. Tiulin (45) has suggested that sieving in benzene or xylene achieves similar results to dry sieving. Undoubtedly, direct sieving of soils in the field has considerable merit under certain definite experimental conditions. It is questionable, however, if such a technique will provide the desired information on soil aggregation where the water stability of the secondary particles is an important factor.

The wet-sieving technique of Tiulin is the best known of the earlier endeavors to find a measure of soil aggregation. According to his method, the soil is slowly wetted by capillarity for 30 minutes and is then transferred onto a nest of sieves immersed in water. The sieves are slowly raised and lowered in the water 30 times. The weight of soil on each sieve is then determined. The bottom sieve in this nest has an opening of 0.25 mm. This procedure has been modified by numerous investigators. Yoder (52), for example, has been one of several who has adapted mechanical methods to the raising and lowering of the sieves. Meyer and Rennenkampff (32) have proposed an automatic siphon arrangement for allowing the water to rise slowly to the top of the nest of sieves and then siphon away rapidly. Pigulevsky (40) has observed that after about sixty repeated immersions most soils continue to yield a constant weight of soil through any given sieve. This amount, which is considered to arise from the mechanical abrasion due to sieving, is applied as a correction factor at the end of the experiment. The wet-sieving technique is well adapted to the separation of large aggregates. It can be used down to a particle size that is larger than 0.1 mm., although 0.25 mm. is more satisfactory as the lower limit.

Elutriation may be used for separating aggregates with diameters between 1 mm. and 0.02 mm. It is particularly useful for making separations below the limit where wet sieving cannot be employed. Several investigators (7, 13) have successfully used the elutriator for making an aggregate analysis. Some were interested in determining the amount of relatively small aggregates in the soil, in order to ascertain causes for the genesis of the aggregate. The use of the elutriator has been criticized on the basis that the density of the

aggregates is not constant and, consequently, that Stokes' law cannot be used satisfactorily for calculating the speed of elutriation. Undoubtedly, this is true to a certain extent. However, there is no reliable technique for measuring aggregates smaller than 0.1 mm. that does not require the use of Stokes' law for determining hydraulic or settling velocities. Another repeated criticism of elutriation, which does not seem justified in light of all the information on the stability of the aggregate, is that the water passing through the elutriator washes out flocculating electrolytes and causes a dispersion of the aggregate. This may be true in certain soils with high salt contents but is not likely to prove serious with most soils. The most serious objections to the elutriator are the difficulties of operation that were discussed in Chapter II.

Sedimentation methods have been used to determine the aggregate distribution in the finer fractions that cannot be separated by sieving. The pipette and hydrometer techniques have been used. Cole and Edlefsen (12) have criticized wet sieving and elutriation on the basis that the action of water disperses many of the aggregates. There may be certain soils in which it is essential to know the extent of this type of unstable aggregation. Cole and Edlefsen have designed a large sedimentation tube in which the particles fall in still water. At the desired time, the tube is laid horizontally and the aggregates are caught on the wall of the tube. A celluloid sleeve is placed inside the tube to expedite the grouping of the various aggregates according to size. Two somewhat important difficulties exist in the use of sedimentation for obtaining a complete curve of the distribution of the various aggregates from their settling velocities. The first has been mentioned in connection with elutriation, namely, the varying density of the aggregates. This is especially true with the larger secondary particles. The other is the possibility of flocculation during sedimentation, particularly the orthokinetic type in which flocculation is accelerated by the downward motion of the larger aggregates.

PREPARATION OF THE SAMPLE. The greatest difficulty in the present methods of determining the distribution of aggregates lies in the manner in which the sample is prepared for analysis. The question of wetting the sample needs considerable clarification. There is no general agreement as to the most satisfactory technique for wetting dried samples. Several investigators have used soils at their field-moisture contents. All large lumps, however, were gently crushed to pass a 2- or 4-mm. screen. It is obvious, however, that this moisture content is a variable quantity. Yoder (52) and Russell (40) have observed that air drying decreases the percentage of large aggregates

in favor of the smaller. This effect is greater the more intense the drying process. The more rapidly the soils are wetted, the greater is the breaking up of the larger aggregates. Immersing the soil in water causes more destruction of the larger aggregates than wetting by capillarity. Spraying water onto the aggregates with an atomizer produces the least destruction of either of these three methods. This degradation of aggregate size on wetting has been shown to be related to the shattering effect of air that is displaced during wetting. Russell states that wetting in a vacuum reduces this shattering to a minimum and offers a possibility of giving a good measure of water-stable aggregates. This does not obviate the disintegration effects of swelling that occur when a dried clod or aggregate is wetted.

The false picture that is so often obtained from aggregate analyses of dried clods from coherent soils, especially from subsurface horizons that rarely become completely dry under natural conditions, is somewhat disconcerting in our present studies of soil structure. Under normal moisture contents, there may be little evidence of a high degree of aggregation. However, when the coherent clods are brought into the laboratory, dried and then wetted by any means, they slake into a large number of fragments. An aggregate analysis shows that such a layer is highly aggregated, which is far from being correct. It is admitted that this effect may not be serious in exposed soils, since the rapid wetting during rainstorms undoubtedly causes a shattering of the aggregates. The impact of the raindrops produces a further dispersive action. Nevertheless, for a large number of soils, it seems plausible to suggest that samples should not be completely dried before making an aggregate analysis, if a true picture of the structure capacity of the soil is desired. It is as important to know whether or not a certain soil will break up into aggregates under natural conditions as it is to find out the percentage of secondary particles that results from a natural breaking up of the soil into aggregates. More experimental results are necessary before this question can be adequately answered.

There is some difference of opinion concerning the desirability of a slight amount of shaking of the sample prior to determining the distribution of the aggregates. Demolon and Henin and others visualize the natural field structure of soils to be constituted of rather stable aggregates. Consequently, a slight preliminary shaking of the sample precedes the analysis. Ten minutes of shaking in a reciprocating shaker have been suggested (7). (It is now believed that this should be reduced to 5 minutes.) Demolon and Henin have recommended shaking for 30 minutes in an end-over-end shaker in a 0.1 per cent calcium nitrate solution; the salt solution is used to prevent defloccula-

tion. The question of shaking preparatory to an aggregate analysis is not as serious as that of wetting. A small amount of shaking, however, seems to permit closer duplication of results than no shaking.

EXPRESSION OF RESULTS. The usual practice in calculating the state of aggregation of the soil is to make an analysis of the sample, with and without dispersion. The total weight of a given fraction in the undispersed state minus the weight of the same fraction when completely dispersed is taken as the quantity of aggregates in the soil having that particular size. For example, if the wet-sieving technique is used, the aggregated sample is passed through the nest of sieves; the weight of material on each screen represents the aggregates plus mechanical separates of this size. Then, the completely dispersed sample is passed through to determine the weight of the mechanical separates in each size range. The percentage aggregation is obtained from the difference between the two distribution curves. It should be mentioned, however, that a certain error is introduced into the calculations with this technique. The mechanical separates that remain on a given screen after the sample is completely dispersed may or may not be the same particles that were associated with the aggregates on this same screen when the sample was not dispersed. It would seem more highly probable that many of them were originally a part of coarser aggregates, and were carried down from coarser sieves after the aggregates were dispersed. The same effect may be obtained in the use of the elutriator. This error may easily be remedied if, instead of attempting to determine a size distribution of the completely dispersed sample, the aggregates on each screen (or in each tube of the elutriator) are dispersed and the material passed through the same sieve (or elutriating tube). The difference in weight before and after dispersion will more nearly give the amount of aggregates of that particular size. Even then the possibility exists that a portion of the resulting mechanical separates may have been associated with the original aggregates. The aforementioned error cannot be corrected similarly in sedimentation procedures of aggregate analysis, since a complete separation of the various sizes of aggregates is not obtained.

Tiulin has suggested that only those aggregates that are larger than 0.25 mm. are responsible for stable soil structure. The percentage of aggregates larger than 0.05 mm. has been used to characterize the "state of aggregation" of the soil. This lower limit was chosen on the basis of the fact that the aggregate-analysis and mechanical-analysis curves cross near this point, which makes it impossible to determine aggregates smaller than 0.05 mm. from the two curves. Some investi-

gators have employed sizes ranging from 0.002 to 0.1 mm. in order to express the percentage aggregation of the soil.

Middleton (33) has suggested the dispersion ratio as a measure of aggregation; this ratio represents the percentage of particles smaller than 0.05 mm. in the aggregated sample divided by the percentage of particles of the same size in the dispersed sample. The lower this ratio the greater is the percentage aggregation of the silt and clay. Vageler and Alten (48) and others have employed the same principle but have used 0.002 mm. as the size limit. It has been recommended that the percentage of mechanical separates smaller than a given size, either smaller than 0.05 or 0.1 mm., which are aggregated into secondary particles larger than this size, be used to express the "degree of aggregation" of the soil. In this way one can distinguish between the total percentage of aggregates or "state of aggregation" and the "degree of aggregation" of the finer mechanical separates.

Other methods of expressing the structure capacity of the soil include the use of certain physicochemical properties along with the percentage of aggregates. Most of these expressions aim at the same objective, namely, to characterize soil aggregation on the basis of how many aggregates are present and to what extent the silt and clay are aggregated.

Porosity Determinations. GENERAL CONSIDERATIONS. Most soil-porosity measurements are based upon determinations of the apparent specific gravity or volume weight of the soil at some arbitrary moisture content. The total porosity is calculated from the real and apparent specific gravity of the soil. At present, many investigators are following the suggestions of Doiarenko and others that the soil-pore space should be divided into capillary and non-capillary porosity. A sample of soil is allowed to reach capillary saturation, which supposedly fills only the small pores with water. The total, capillary and non-capillary porosity may be calculated from the volume, weight, moisture content and density of the soil particles. The capillary porosity is equivalent to the moisture content in per cent by volume. The non-capillary porosity is the difference between the total and capillary porosity.

The density of water in these calculations is generally considered to be equal to that of free water. This is not true since the adsorbed water has a higher density than free water. Russell (40) suggests that too large a value for total porosity will be obtained if the density of the particles is determined in water; also, the capillary porosity will be too high if the density of the soil water is considered equal to that of free water. These two overestimates counterbalance each other to give a fairly accurate measure of the non-capillary or air capacity.

He advocates that the density of the particles be determined in a non polar liquid and that the volume of the pores filled with air be measured directly in an ordinary Boyle's law apparatus, as devised by Torstensson and Eriksson (47).

Aside from this criticism of the usual technique for determining porosity, there is always the question of the swelling effect of water on soil porosity. When a dry soil is wetted, its total porosity is increased because of swelling. This influence is considered to be serious by some investigators. It is the author's opinion, however, that the swollen soil represents the desired condition in most instances, particularly in the humid region. With the possible exception of the surface horizon, the structural properties of a soil profile should be evaluated in the moist condition. It seems feasible, therefore, to choose an appropriate time, as far as moisture is concerned, for collecting field samples for pore-space determinations.

Perhaps the most serious objection to present methods for measuring porosity is the complete lack of uniformity in designating the tension at which the capillary pores are saturated. Some investigators place the soil in contact with a free-water surface and allow the soil to saturate by capillarity; others use the moisture content at field capacity when free water no longer drains from the soil; others have placed the soil on a layer of sand at a distance of several centimeters above the water level. The most recent trend, however, is to measure the moisture content at various tensions to obtain a picture of the distribution of the various sizes of soil pores. This technique will be discussed on page 223 of Chapter VI.

VOLUME-WEIGHT DETERMINATIONS OF NATURAL SOILS. Volume-weight determinations require either a core of soil in its natural structure or a clod that represents natural breakage. The obtaining of core samples is relatively simple if no roots or stones occur in the soil. The core sampler is pushed or driven into the soil to the desired depth and then removed. Since the volume of the sampler is known, it is easy to obtain the volume weight, air and water capacity of the core of soil at any desired moisture content. Many samplers are provided with a brass or celluloid casing which holds the core and permits easy removal and handling of the sample during wetting and weighing.

There are several factors that should be considered in obtaining core samples by this technique. The forcing of the sampler into the soil may shatter or compress the soil, depending upon the moisture content and the method of sampling. Hammering tends to cause shattering and rapid pushing produces compression. The size of the sampler affects the amount of compression. Narrow sampling tubes

favor compression; a 3- to 4-inch diameter is a satisfactory and convenient size for most work. A typical example of such a core sampler is shown in Figure 37a. Very slight effects on the natural soil structure are produced by hammering this instrument into the soil. A wooden plank on top of the heavy steel ring aids in absorbing the shock of the hammer. Many soil cores may be extracted by having a large number of brass inner rings. After the sample is removed from the soil, the core is placed in a paraffined, pint-size, ice-cream container and sealed. In this way the cores can be stored until com-

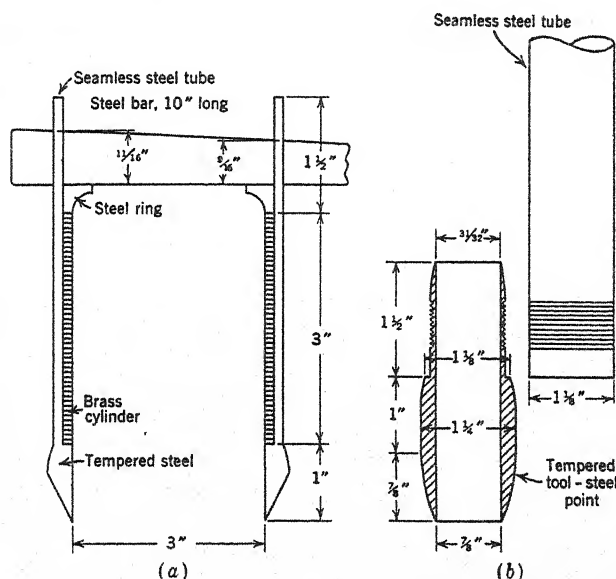


FIG. 37. Soil-sampling tubes of Bradfield (a) and Veihmeyer (b).

plete porosity measurements can be made, without serious moisture losses.

Veihmeyer (49) has perfected a modification of the usual King type of core sampler. The diameter of the cutting edge is slightly less than that of the tube so that the soil slides through the tube easily. There is also a small enlargement on the outside of the tool above the cutting edge to make the tool more easily extracted from the soil, especially when deep core samples are taken. This sampler has the distinct advantage of making deep sampling possible. (See Figure 37b.)

Attempts have been made to determine the volume weight by boring a hole of known volume into the soil and filling it with sand or

some liquid that will not penetrate the walls of the hole. A method of this type will permit measurements of total porosity but is not easily applicable to pore-space distribution determinations.

Sekera (42) introduced the idea of determining the porosity and water capacity of soils by using natural clods. After the clod is allowed to saturate capillarily with water under a tension of about 7 cm., its volume is determined by measuring the displacement of the wet clod in mineral oil. If the wet clod is then removed from the oil, washed with xylene until free from oil, dried and weighed, it is possible to determine the air capacity. This method is well adapted to the study of the porosity of the larger units into which the soil crumbles. It may give an erroneous concept of the porosity of the profile as a whole if there are considerable pores between these aggregates or fragments. In order to prevent slaking, it is essential not to dry the clods before saturation by capillarity. The total porosity of the clod may also be determined by coating with a thin layer of paraffine and weighing in both air and water. The volume is calculated from the amount of water displaced.

It might be concluded that pore-space measurements to determine the air and water capacities of a soil are important phases of any study on soil structure. Soil aeration and water movement are intimately associated with the amount and size of the pores. Accurate samples in the natural state are the first requisites of successful porosity determinations. In most instances, the investigator should exercise care in choosing a suitable moisture content for obtaining such samples. When the soil is too dry or too wet, a sample that represents the normal structure will be the exception rather than the rule.

Permeability Determinations. The permeability of the soil for water or air is a function of the amount and size distribution of the pores. It is an index of certain structural relationships. Inasmuch as permeability measurements cannot be divorced from a discussion of water movement in soils, this method of evaluating soil structure will be considered in Chapter VI.

CULTURAL CHANGES AFFECTING SOIL STRUCTURE

Every farmer who has cleared virgin timberland or plowed virgin prairie sod is clearly aware of the fact that the soil is not as loose and friable after continued cropping and cultivation as when it was first plowed. Even now in areas of extensive cultivation, one can find small tracts of nearly virgin timber or prairie that may be used as yardsticks of the extent of deterioration of structure during farming operations.

Numerous examples have appeared in the literature which show that cultivated soils have 30 to 40 per cent less organic matter and nitrogen than adjacent virgin areas. Common observation has shown that soil tilth has become poorer along with the decreases in humus and nitrogen.

Cultivation of soils in this sense is generally considered a synonym of soil culture. It means that the land has been plowed, planted to crops and tilled. Most of these agricultural processes lead to degradation influences upon soil fertility and soil structure. Some may have a regeneration effect. Jenny (23) has shown that sixty years of cultivation of the Putnam silt loam in Missouri has led to a 38 per cent decrease in organic matter, a 33 per cent decrease in available bases and a corresponding diminishing of aggregation.

TABLE 21
EFFECT OF CULTIVATION ON THE AGGREGATION OF SOILS

Soil type	Carbon	State of aggregation		Degree of aggregation	
		> 0.1 mm.	> 0.05 mm.	> 0.1 mm.	> 0.05 mm.
	per cent	per cent	per cent	per cent	per cent
Lansdale silt loam <i>F</i> ...	5.25	38.8	44.3	47.1	55.5
<i>C</i> ...	0.32	10.8	19.3	11.0	21.4
Lansdale silt loam <i>F</i> ...	2.20	43.4	44.4	45.4	49.4
<i>C</i> ...	1.40	16.6	27.1	17.5	30.5
Chester silt loam <i>F</i>	3.22	28.0	37.2	42.2	54.9
<i>C</i>	1.99	14.5	28.7	19.0	36.6
Chester silt loam <i>F</i>	3.47	38.2	38.0	46.2	50.8
<i>C</i>	1.08	12.0	22.4	14.1	28.4
Putnam silt loam <i>P</i>	2.13	27.8	44.5	28.8	47.8
<i>C</i>	1.31	6.8	13.0	7.0	14.2

* *F*, *P* and *C* represent virgin forest, virgin prairie and cultivated samples, respectively.

It has been shown (Table 21) that the state and degree of aggregation of cultivated forest and prairie soils are much lower than of soils from adjacent virgin areas. Not only are there fewer total aggregates present in the cultivated soil, particularly the larger granules, but there is also a lower percentage of fine particles in the aggregated

state. Degradation of aggregates is highly correlated with the decrease in organic matter.

Bradfield (9) has reported that the porosity of a clay soil in northwestern Ohio has decreased from 16 to 18 per cent below that of relatively virgin forest, during 40 years of farming to intertilled crops. A porosity of 60.3 per cent in the surface foot of a soil, producing its first corn crop after the sod had been plowed, gave a corn yield of 80 bushels per acre as compared with 20 bushels in the cultivated field across the fence, where the soil had a total porosity of 50.5 per cent. This difference in yield is not primarily due to fertility since fertilization gives little response on this soil unless its physical properties are first improved.

These technical data confirm the common observation that tillage of soils deteriorates structure as well as fertility. This deterioration apparently is expedited by the fact that the structure-improving qualities of virgin plant associations are completely removed by the plowing operation. The summation of the effects of decreased organic-matter production, increased organic-matter decomposition, increased leaching, the impact of raindrops on exposed soil and the mechanical manipulation of tillage implements is a deteriorated structure. From an agricultural point of view, it is necessary to evaluate the various means by which structure is broken down during soil culture as well as the methods that may be used to regenerate a favorable structure.

Crop Effects. The growing crop affects the structure of the soil, directly and indirectly. The direct effects may be divided into two categories: (1) the protection afforded by the leaves and stems against the impact of raindrops, which retards the deterioration of structure, and (2) the development of granulation and porosity through root activity, which aids in the regeneration of structure. The indirect effects may be listed as those changes in granulation that are caused by the organic matter produced by plant growth. They have been thoroughly discussed earlier in this chapter.

CANOPY PROTECTION. At present there is little recent information on the extent to which a canopy of vegetation preserves the structure of the surface soil by preventing dispersion of the soil that would result from the direct impact of raindrops. In 1874, however, Wollny (51) observed that rye, peas and vetch protected the soil from the raindrops to such an extent that the non-capillary porosity of a shaded, humus-containing, calcareous sandy soil was 34 to 53 per cent higher than adjacent unprotected soil. A part of Wollny's results are shown in Table 22. It is seen that the protective covering of the above-ground portion of plants is very effective in preserving the content of

TABLE 22

EFFECT OF PLANT COVER UPON SOIL POROSITY (WOLLNY, 51)

Soil type	Type of cover	Total porosity		Non-capillary porosity	
		Cover, per cent	No cover, per cent	Cover, per cent	No cover, per cent
Humus calcareous sand	Rye	62.2	58.3	33.5	22.1
	Peas	63.6	59.2	32.3	21.0
	Vetch	69.2	65.5	39.0	29.0
Quartz sand.....	Sweet clover	46.0	45.4	43.9	34.9
Loam.....	Sweet clover	55.4	50.9	45.3	31.9
		Number of plants per 1000 sq. cm.		Per cent decrease in volume from May 1 to September 12, 1879	
Humus calcareous sand	Oats	0		12.8	
		3		9.2	
		6		6.9	
		12		6.1	
		24		4.9	
		12 (fertilized)		5.3	

large pores in the soil. It is important to note that the deterioration of structure resulting from the impact of raindrops is reflected more in the decrease in the volume percentage of large pores than in a diminution of total porosity. For example, the total porosity of the unprotected soil averages only about 6 per cent less than that of the protected, for the rye, peas and vetch crops; the non-capillary porosity is about 31 per cent less.

Another significant point in this table is the effect of the number of plants and density of foliage on structural changes. As the number of plants per unit surface area increases, there is a reduction in the magnitude of the soil volume decrease. Moreover, twelve fertilized plants are more effective than twenty-four unfertilized ones. These results simply state that the canopy influence of vegetation is more effective the denser the foliage and the more rapid the rate at which the protective cover is established. The fundamental cause of these

structure changes is the dispersion of the aggregates in the immediate soil surface and the movement of the clay and silt particles into the larger pores. This decreases the content of large pores and increases compaction. Recent experimental data suggest that these structural changes in the immediate surface may often be more important than differences in structure deeper within the soil profile.

ROOT INFLUENCES. It is difficult to separate the direct and indirect influences of root activity on soil structure. One cannot distinguish as yet between the aggregation effects of root pressure, binding qualities of root hairs, the production of organic matter, moisture changes resulting from water usage or any possible root excretions. The relation of plant roots to aggregation has been discussed in connection with the genesis of soil structure. A few comments on porosity changes resulting from root penetration will be helpful, even though few reliable experimental data regarding these changes exist. From the standpoint of increased porosity resulting from old root channels, it has been frequently observed that certain core samples have extraordinarily high percolation rates, owing to these canals through

the core. These old root channels are especially obvious in soils under forest vegetation or under crops having rather large root systems, such as sweet clover and alfalfa. There can be no doubt as to the beneficial effects of roots on soil porosity. The exact magnitude of these influences must await more reliable information.

OBSERVED EFFECTS OF DIFFERENT CROPS. The most outstanding investigations of the influence of growing crops on soil structure have been accomplished by European investigators, particularly the Russians. The reader is referred to the work of Sokolovsky (44) and Williams (50) for enlightening discussions of the relation of

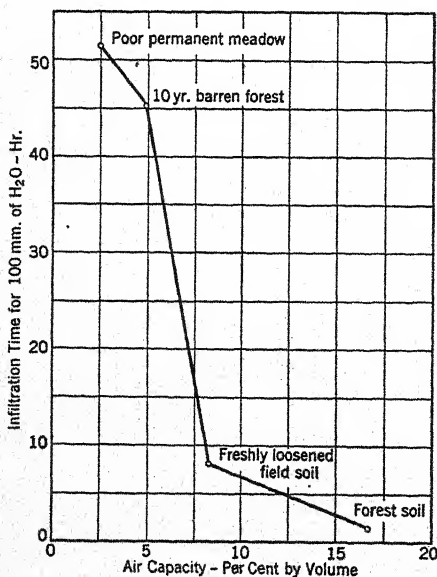


FIG. 38. The relation of infiltration rate to the air capacity of soils (Burger, 10).

vegetation, particularly grasses, to soil properties.

Burger (10) made extensive studies of the effect of forest cover

on the properties of the soil. Some of his results are summarized in Figure 38. This graph is almost self-explanatory and shows that good forest cover produces a high non-capillary porosity, which is reflected in a greater infiltration rate. These results call attention to two facts. First, the rate of water intake by soils is dependent upon the content of large pores. Second, the handling of the land determines the number of non-capillary pores. Doiarenko (44) reported results which showed that grass in a rotation had a great effect upon porosity relationships. The non-capillary porosities of fields in permanent fallow, in flax in rotation without grass and in flax in rotation with grass were 4.9, 5.2 and 19.7 per cent, respectively. It is significant to note that the rotation did not materially affect structure until grass had been included.

The Russian scientists have studied in considerable detail the effect of cultivation and cropping on the deterioration and regeneration of soil aggregation. These

outstanding investigations are undoubtedly the result of the viewpoint of W. R. Williams, who developed the concept that soil structure was the key to fertility and that a grassland system of agriculture was the means of achieving a desirable structure (50). Pavlov (44) found that cropping land to cotton resulted in a rapid breakdown of the soil granules. Clover and alfalfa improved both the micro- and macro-structure of the soil. The detrimental influences of cotton and the beneficial effects of clover on aggregation are shown in Figure 39. It is seen that one year of cotton, following two years of clover, decreased the aggregation from 34.7 to 17.5 per cent; one year of clover after two years of cotton increased the aggregation to 22.8 per cent.

Heltzer (18) has shown that alfalfa increases the organic-matter content and state of aggregation of soils. Some of his more outstanding results are given in Table 23. These results indicate considerable

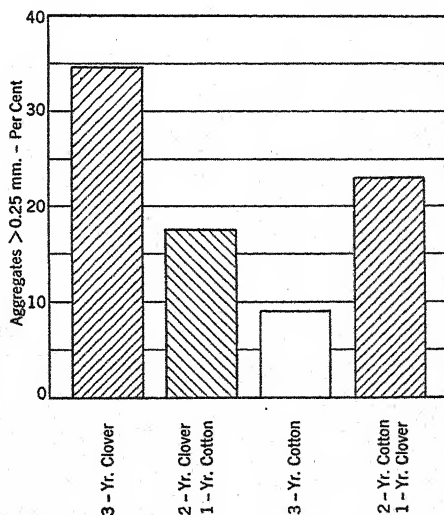


FIG. 39. The influence of intertilled crops on soil aggregation (Pavlov, 44).

beneficial effects of alfalfa. They are confirmed by practical observation in the field.

TABLE 23

THE EFFECT OF COTTON AND ALFALFA UPON SOIL AGGREGATION
(AFTER HELTZER, 18)

Crop	Per cent carbon	Aggregate composition in per cent		
		2-0.25 mm.	0.25-0.1 mm.	Smaller than 0.1 mm.
Field cropped to cotton for a long period.....	0.52	2.23	12.10	85.67
3-year-old alfalfa.....	0.80	15.01	49.51	34.58
5-year-old alfalfa.....	0.92	27.32	47.35	25.33

Ilmenjew (21) studied the effect of various crops and manuring on the state of aggregation of a podsol. A summary of his results is shown in Figure 40. Some very significant facts may be gleaned from this graph. In the first place, it is obvious that those crops which provide more cover and have a more fibrous and extensive root system have a greater effect upon soil granulation. Rye is better than oats because it is on the land longer and makes a bigger growth. In the second place, it is seen that manure increases granulation over that of the check plot. This increase, however, is not as great as the differences between the effects of the various crops.

Rostovzeva and Avaeva (38) investigated the role of perennial grasses in the formation of stable structure of soils and found that deteriorated structure can be regenerated by means of grass vegetation. A summary of their investigations is shown in Figure 41. These data point out that different types of soil vary greatly in their stability of structure under cultivation. For example, the gray forest soil suffered severe deterioration of structure under cultivation. The chernozem, a prairie soil, suffered least. A grass podsol had a much more favorable structure than a forest podsol. Cropping these cultivated lands to perennial grasses restored a large percentage of the original granulation. Two years of grass produced an 81 per cent increase in the aggregation of a cultivated gray forest soil, a 65 per cent increase in the granulation of a cultivated grass (turf) podsol and a 37 per cent increase in the aggregation of a degraded chernozem. These data confirm the concept of Williams that stable soil structure

is the product of grass vegetation. Similar tendencies have been obtained by the author on numerous occasions, but none of his data approaches the magnitude of the effects found by the Russians.

Effect of Tillage. Tillage may or may not be beneficial to soil tilth, depending upon the moisture content at which the soil is worked and the number and types of operations. There is a growing belief that any beneficial effects that may be produced by one tillage

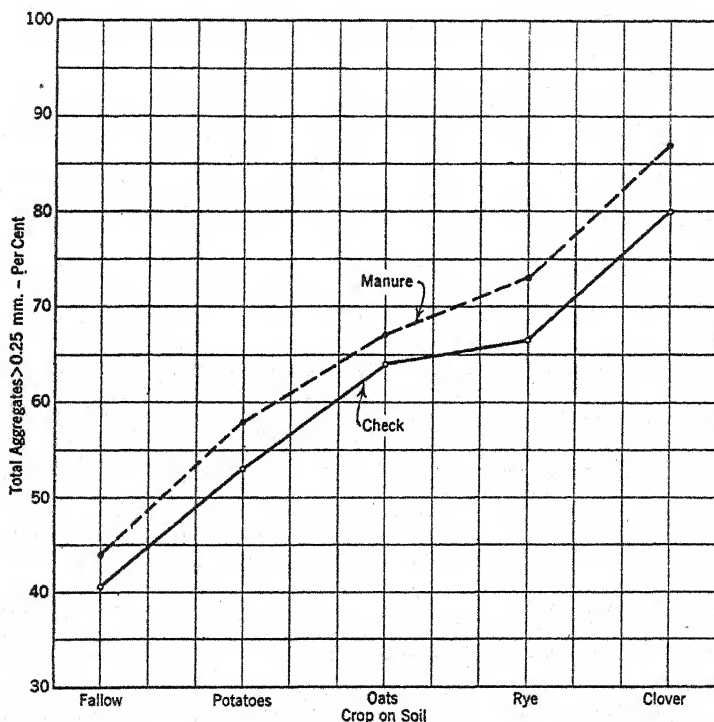


FIG. 40. Soil aggregation under various cropping systems (Ilmenjew, 21).

implement are often nullified by excessive subsequent operations. One of the best surveys to date on the effect of tillage on soil structure is found in the paper of Russell (40). This paper points out that practically all information in this field is confined to Russian investigators. Fortunately, Russell has had access to this work. Most of the remarks that follow refer to his review of existing data; the original references may be found in the comprehensive bibliography of the original paper.

In the discussion on soil consistency, the importance of the soil moisture range in tillage operations was strongly emphasized. At-

tention was called to the fact that the mechanical manipulation of the soil in either the plastic or hard consistency ranges will usually result in unfavorable physical conditions. Vilensky and Germanova

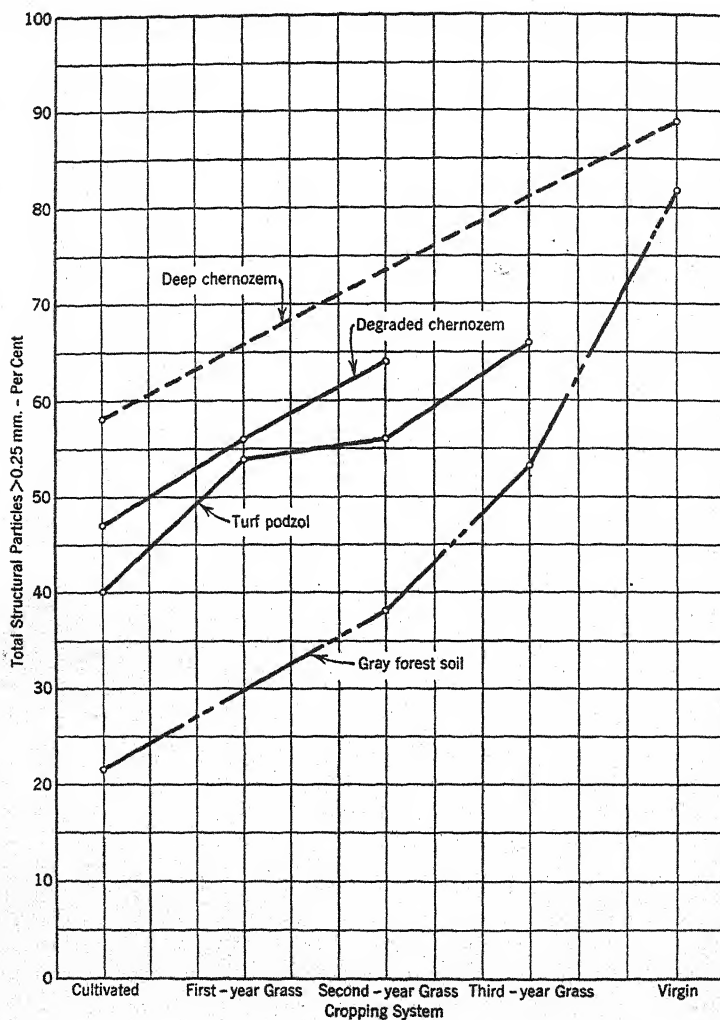


FIG. 41. The influence of grass crops on stable soil aggregation (Rostovzeva and Ayaeva, 38).

have found that there is an optimum moisture content for structure formation. At low moisture contents particles do not stick together and at higher moisture percentages the soil dries into hard clods.

Russell suggests that this optimum moisture content is probably close to that of the sticky point where most of the pores are filled with water and close packing between particles can occur. There is reason to believe, however, that this value for plastic soil lies slightly below the moisture content where the soil becomes plastic. This opinion is based upon observations in the field and experimental data on the soil dynamic factors involved in tillage. The Russian work is based upon the aggregation of a loose, fine soil. It is probably necessary to exert a slight puddling or compression action on loose soil to cause it to stick together.

A different situation exists in compacted field soils. If they are cultivated too dry, large clods or fine dust are produced, depending upon the texture. If they are tilled too wet, puddling causes the formation of clods. Such a soil must be manipulated at the moisture content of friable consistency. This fact is illustrated to a certain extent by the data of Tsyganov, who compared the granulation produced when the soil is cultivated near its wilting point and near its field capacity. A brief summary of his results is given in Table 24. Plow-

TABLE 24
THE EFFECT OF TILLAGE ON SOIL AGGREGATION (TSYGANOV, 40)

Type of tillage practice	Moisture content at cultivation	Percentage of water-stable aggregates	
		Larger than 5 mm.	Smaller than 0.25 mm.
Before plowing.....	near wilting point	48	22
	near field capacity	13	40
After 2 plowings and 4 diskings...	near wilting point	54	19
	near field capacity	44	24
After 2 plowings and 20 diskings.	near wilting point	39	28
	near field capacity	30	31

ing and discing near the wilting point caused a decrease in aggregation. Moderate discing at the field capacity gave the maximum amount of aggregates. Other Russian investigators have observed similar effects.

Nikrassov studied the effect of plowing, cultivating, harrowing, rolling and hoeing on the capillary and non-capillary porosity of the soil. The major influence of tillage was on the non-capillary porosity.

A compacted soil is loosened by cultivation which increases the content of large pores in the immediate surface. This increase in non-capillary porosity benefits aeration and water absorption.

It seems clear that tillage operations may have varied effects upon soil structure depending upon the nature of the implement and the moisture content of manipulation. It is believed that any degradation of soil structure that results from tillage is not so much a question of unfavorable factors resulting from plowing as from the excessive manipulation of the soil after plowing. The present interest of agricultural engineering in the whole soil-tillth problem foreshadows valuable information in the future on tillage and soil structure changes. These relationships will be discussed in more detail in Chapter IX.

The Effects of Fertilizer, Manure and Lime. Manure and lime have been looked upon by farmers and scientists as the practical solution of soil-structure problems. As far as lime is concerned, attention has been previously called to the fact that field and laboratory data do not confirm any direct effect of lime on soil structure. It is recognized, however, that liming promotes greater development of vegetation and production of organic matter, which usually causes a regeneration of structure. It seems safe to conclude, in light of all existing information, that liming promotes better soil structure through its indirect effects upon organic-matter production and microbiological activity.

The effects of manure on soil structure are far from being clearly understood. Like lime, manure seems to exert some physical influence upon the soil that is reflected in good crop production. C. E. Thorne, the founder of the Ohio Agricultural Experiment Station, after analyzing all the manure plots at this station, came to the conclusion that the beneficial effects of manure are essentially the result of the fertility elements contained therein. This conclusion has been challenged by many agronomists on the basis that the naturally favorable physical condition of the soil at the Ohio Station was responsible for the data upon which Thorne based his conclusions. The experimental facts on other soils, however, seem to point to the same conclusion as Thorne's, with some reservations.

The data of Ilmenjew in Figure 40, for example, indicate that manure has some beneficial effects on soil aggregation, although the effects of manure are not as great as the influences of different crops. Bertramson and Rhoades (8) have reported that normal applications of manure show no effect upon the state of aggregation or porosity of the Marshall silty clay loam, over a fifteen-year period. Musgrave and Norton (34), on the other hand, have found that annual additions of 8 and 16 tons of manure increased the infiltration rate of the Mar-

shall silt loam and Shelby loam. This effect is attributed to an increase in porosity of the surface layer. A significant, although not large, increase in the granulation of the Putnam silt loam has been observed on the fertility plots at the Missouri Agricultural Experiment Station. The experiences of Russian investigators relative to the effects of manure on soil structure vary. On a podsolized loam, manure caused both increases and decreases in aggregation. On a degraded chernozem, it exhibited little influence on aggregation. Sokolovsky (44) has recognized the complex relationships that are involved upon manure applications to the soil. He has visualized the possibility of a dispersive action on soil aggregates, owing to the ammonia produced during decomposition and to the protective action of organic colloids on the mineral fractions. On the other hand, aggregation may result from the binding effects of the humus produced in its decomposition. Its influences are considered to be dependent somewhat upon the season and the type of soil.

In general, it appears that manure exerts a favorable effect upon granulation and aeration in the soil but that this effect is not very permanent. With the possible exception of heavy applications, the fertility factor usually outweighs the physical. Applications of manure to sods will increase the growth of these structure-improving crops and benefit structure in this way.

Little is known concerning the effects of fertilizer on soil structure. Theoretically, large applications of sodium salts should cause a dispersion of the soil aggregates. Such a degeneration of structure will probably not occur under normal agricultural practices. Increased foliage and root production as a result of fertilizer applications undoubtedly have a large influence on the preservation and partial regeneration of structure. This should be especially true on sod crops. In summing up the possible effects of fertilizers, manure and lime on soil structure, it should be emphasized that they will usually benefit the sod crop in the rotation to such an extent that outstanding increases in soil tilth will often occur.

Effects of Drainage. The primary influence of adequate drainage on soil structure lies in the removal of excess water, with a corresponding increase in the air capacity. This increase in aeration is accompanied by greater root development, more intense bacterial activity and the promotion of oxidation processes. The combined effects of these factors lead to better granulation. In most cases, therefore, drainage may be visualized as a positive structure-forming factor. A more effective non-capillary porosity and stable aggregation follow in the wake of a lower degree of saturation with water.

Effects of Irrigation. Sokolovsky (44) has emphasized the fact that the problem of the formation of good structure requires special consideration under irrigation. The breakdown of aggregation in the surface during irrigation leads to crust formation, which produces unfavorable air and water relations for plant growth. The water that is added is thereby rendered partially ineffective for crop production. The economic importance of this crusting effect is well illustrated in the words of Sokolovsky, who says: "Here also extremely unfavorable physical properties of the soil draw our attention, properties which make tilling hard and difficult, entail considerable waste of water due to surface drying (the crust): All this heavily affects the plant, for it always experiences a deficit in water; the seeds, when germinating, suffer from lack of air and the seedlings force their way through the crust with difficulty; the crust, when cracking, cuts the delicate stalks as with a knife; water, so precious in these regions, is to be greatly wasted without economy."

Cole (11) has observed that irrigation of several California soils by the basin method, where the whole surface is covered with water, causes a puddling of the surface. This puddling action increases the cloddiness of the soil and obviates the beneficial effects of any tillage operations prior to irrigation. In general, the finer the condition of the soil surface before irrigation, the coarser are the clods after the water is applied.

It appears, therefore, that the effect of irrigation water in the arid regions on the structure of the surface soil is similar to that of natural rain in the humid regions, unless water-stable aggregates exist in the surface layer; water causes dispersion of the secondary particles and a decrease in non-capillary porosity follows. Alfalfa and grass sods have the same desirable effects under irrigation as they have in humid regions.

AGRICULTURAL SIGNIFICANCE OF STRUCTURE

Soil structure is the key to soil fertility! This concept has been recognized by eminent soil scientists for a long time. Hilgard (20) stated that one must first determine the proper physical conditions in the soil and subsoil before judging the natural fertility of the soil in terms of plant growth. Wollny maintained that the response of various crops to fertilization depended upon a favorable soil structure. Williams of Russia developed an extensive grassland system of agriculture on the thesis of producing and maintaining good structure. These pronouncements, along with many others of more recent

origin, apparently have been heard by farmers, agriculturists and some scientists only as faint whispers among the arguments for a chemical solution of profitable crop production.

Common observations as well as certain experimental results have shown that some soils do not respond to fertilization even though they are not highly productive. Crop yields vary between wide limits depending upon the amount and distribution of rainfall. Chemical analyses show that sufficient fertility elements are present to produce abundant crops. What, then, are the factors that limit the anticipated crop production?

It is known that plants require nutrients, water and air for growth. The amount of nutrients in the soil is usually taken as an index of fertility. The air and water relationships are dependent upon structure. If inadequate moisture is present, the plant cannot utilize nutrients; it cannot carry on its normal physiological functions. If excessive water or insufficient air (oxygen) is present, the plant cannot make efficient use of the nutrients in the soil. The growth of plant roots and the germination of seeds require favorable conditions for respiration. If there is a limited supply of oxygen within the soil as a result of poor structural conditions, respiration processes are hindered; germination and growth are retarded. Moreover, a small root system restricts the soil volume in which nutrients are available to the plant. Consequently, a low air capacity may affect plant development in more than one way.

In addition to the effect of improper air and water relationships on the germination and growth of plants, bacterial activity is also hindered. Nitrification processes cannot proceed normally and the production of nitrates from organic matter is diminished to the point that the plant suffers. Other aerobic biological activities are impeded.

These facts indicate that abundant nutrients in the soil do not insure good crop production. The investigations of numerous workers emphasize that insufficient attention has been given to providing a favorable environment for the germination of seeds and growth of crops. Little consideration has been paid to the development and fertilization of cropping systems from the standpoint of their effects upon structure and other physical properties of soil. The future promises a rapid growth in the interest and emphasis that will be placed upon the agricultural value of good soil structure.

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CHAPTER VI

SOIL WATER

CONCEPTS OF SOIL WATER

Soil-structure studies have shown that the soil is a porous medium which is made up of various sizes of pores. Water that enters the soil either remains in the pores or percolates through them to lower depths. The character of the soil pore space determines to a considerable extent the disposition of infiltrated water. Water retention and movement, therefore, constitute two important phases of soil-moisture relationships. It is realized that a discussion of soil water may be approached in several ways. In light of the importance of this subject in many soil-physics problems, however, it seems advisable to acquaint the reader first with the various concepts of soil water and then to present the different ideas on soil-moisture movement. The first will include the more important attempts to characterize the way that water is held in soils and to formulate so-called equilibrium points for expressing differences in the water relations of soils. The second will involve the movement of infiltrated water, in both the liquid and vapor phases.

Two slightly different viewpoints exist in the literature relative to soil-water relationships. The earlier concepts were primarily based upon the so-called capillary-tube hypothesis, which considered the soil to be constituted of numerous capillaries of varying dimensions. Water retention was viewed as a function of the tension of water films around particles. The contributions of L. J. Briggs and his associates from the United States Bureau of Soils to this school of thought are well-known to all soil students. It is interesting that another representative of the Bureau of Soils, E. Buckingham, introduced the concept of using energy relations to characterize soil-moisture phenomena. The development of this idea by Gardner, Haines and others has caused a gradual displacement of the old capillary-tube hypothesis by this new concept. Essentially, there is no conflict between the two concepts. In the discussions to follow attention will be called to the various modifications of the older concepts as well as to the more modern theories. This will enable one to picture the development of our ideas on soil

moisture and will permit a clearer understanding of the energy relationships involved in soil-water problems.

The Capillary-tube Hypothesis. Briggs (7), in 1897, in trying to develop a concept of the mechanics of soil moisture, conceived the idea that capillary water existed as a continuous and tightly stretched film around the particles. The forces arising from the curvature of these capillary-water surfaces were visualized as the principal cause for the retention of soil moisture. Naturally, this retention would then be dependent upon the number and size of the capillary spaces. According to this concept, water movement takes place from the thicker to the thinner films and the rate of movement is related to the difference in curvature of the films, the surface tension and the viscosity of the liquid.

Briggs proposed that soil water should be classified into three kinds:

1. **HYGROSCOPIC WATER**, which is adsorbed from an atmosphere of water vapor as a result of attractive forces in the surface of the particles.
2. **CAPILLARY WATER**, which is held by surface tension forces as a continuous film around the particles and in the capillary spaces.
3. **GRAVITATIONAL WATER**, which is not held by the soil but drains under the influence of gravity.

Subsequent investigations by Briggs and his associates as well as by other workers seemed to concentrate on an analysis of capillary water in terms of its availability to the plant, especially from the points of view of the optimum and minimum moisture contents for plant growth. Briggs and McLane (8) developed the moisture equivalent as an expression of the ability of a soil to hold water under a centrifugal force 1000 times that of gravity. This value was considered to represent the moisture in the smaller capillary pores. Later, Briggs and Shantz (9) introduced the idea of using the wilting coefficient to express that moisture content of the soil at which plants permanently wilt. Wilting was thought to occur when the soil could not release water to the plant at a rate sufficient to offset transpiration. Widtsoe and McLaughlin (62) differentiated three phases of capillary water on the basis of their experiments on the movement of water in irrigated fields. These were (1) the maximum capillary-water capacity or the water content at capillary saturation against gravity, (2) the optimum capillary capacity in the moisture range that is most favorable for plant growth, that is, the field-water capacity, and (3) the

"lento-capillary point," or the point where moisture movement is very slow. According to their concept, root hairs could draw water from the soil at lower moisture contents only by being in direct contact with a moisture film. The moisture content of the lento-capillary point is nearly twice as great as that at the hygroscopic coefficient. They have stated that it cannot be defined with precision.

Bouyoucos (5, 6) suggested a new classification of soil water on the basis of results obtained from freezing point and dilatometer studies. The main feature of this classification was the introduction of the idea of "unfree" water—water which is held so tightly to the colloidal particles that it is not readily frozen and is not readily available to plants. The classification of Bouyoucos may be summarized as follows:

1. GRAVITATIONAL WATER—moves under force of gravity; is super-available to plants.
2. FREE WATER—freezes at -1.5°C. ; is available to plants.
3. UNFREE WATER—does not function as a solvent.
 - a. *Capillary*—freezes at -4°C. ; is very strongly adsorbed; slightly available to plants.
 - b. *Combined*—does not freeze at -78°C. ; is non-available to plants.

The "unfree" water was visualized as being held by the soil capillaries either in a state of physical adsorption on the colloids, in loose combination with the colloids or as water in the small capillaries.

The concept of Bouyoucos was criticized by both Keen (25) and Parker (40). The former showed that the amount of "unfree" water varies with the moisture content and suggested that the forms of water proposed by Bouyoucos are approximate equilibrium values only and do not indicate any abrupt change in the physical condition of soil water. Parker proved that the soil does not render this "unfree" water from functioning as a solvent. He also pointed out that the relation of vapor pressure, evaporation, freezing-point depression and the absorbing power of seeds for moisture all show similar phenomena when compared as functions of the moisture content. This similarity should be expected since the various curves represent the force with which water is held at different percentages of water in the soil. It is interesting to note that, even though the curves cited by Parker indicate definite energy relations, he concluded that the Briggs' classification of water into hygroscopic, capillary and gravitational is satisfactory.

In spite of the criticisms of the Bouyoucos concept, the idea called attention to the fact that some soil water is held with great force in

the capillaries, although no attempt was made to explain the observed results on the basis of the capillary network of the soil. Even the arguments that were advanced to refute Bouyoucos contained many of the essential features of the energy concept of soil moisture.

Zunker (64) has classified soil water into the following groups:

1. OSMOTIC WATER—in cells of organic matter (bacteria, etc.).
2. HYGROSCOPIC WATER—amount of water in the soil when it shows no heat of wetting. Attracted on surface of particles by free surface energy forces.
3. CAPILLARY WATER—held by capillary forces in fine soil pores that are connected with the ground water.
4. "HELD" WATER (Haftwasser)—water held by surface tension forces on soil particles, under normal pressures and capable of movement, but not in union with the ground water.
 - a. *Film water*—water on soil particles as a skin.
 - b. *"Pore angle" water*—water held in the angles formed by the points of contact of particles.
 - c. *Capillary "held" water*—water held by capillaries not connected with ground water.
5. GRAVITATIONAL WATER—water found in downward or horizontal movement within the zone of aeration.
 - a. *Capillary gravitational water*—water that moves downward and laterally in the capillary pores by means of gravity and capillarity.
 - b. *Downward gravitational water*—water that moves by gravity through the non-capillary pores to the ground water.
6. GROUND WATER—water that fills the tension-free pore space
7. WATER VAPOR—water in vapor form in the soil pores.

This elaborate scheme of classification is an enlargement of the capillary-tube hypothesis. It is worthy to note, however, that true capillarity is considered only in the case of connection of the pores with a ground-water table. Moreover, the vapor phase of soil water is given some attention.

Lebedeff (31) has suggested a simpler modification of the Briggs' classification by emphasizing the importance of water vapor and film water in soils. The various forms of soil water suggested by him are:

- ✓1. WATER VAPOR—water which is controlled by vapor-pressure equilibria. Movement takes place from higher to lower vapor pressures.

2. **HYGROSCOPIC WATER**—water molecules held on surface of particles by forces of adhesion. Movement occurs in the vapor phase from moist to dry areas.
3. **FILM WATER**—water under the influence of molecular forces of cohesion. Movement takes place slowly as a liquid from thicker to thinner films.
4. **GRAVITATIONAL WATER**—water that moves under force of gravity.

Although Lebedeff's classification is essentially a simpler form of Zunker's proposals, it is important to keep it in mind in connection with soil-water movement. Lebedeff's contributions to the movement of soil water have been the basis for this proposed classification. Like the suggestions of Briggs, Widtsoe and McLaughlin, Bouyoucos and Zunker, it is another example of the capillary-tube point of view.

Energy Relationships in Soil-moisture Retention. THE CAPILLARY POTENTIAL. In 1907, Buckingham (10) introduced the idea that the flow of water through the soil could be compared to the flow of heat through a metal bar or of the flow of electricity through a wire. The driving force, or the cause of the capillary current, was visualized as the difference in attraction for water between two portions of the soil that are not equally moist. He suggested the term "capillary potential," characterized by the Greek letter ψ , to express the value which measures the attraction of the soil at any given point for water. For example, a moist blotter will soak up less water than a dry one. Part of the attractive forces within the blotter have been satisfied by previous moistening. Similarly, it will require less energy to pull a given amount of water from a soil that is nearly saturated with water than from a slightly moist soil. This quantity depends upon the soil and the water content. In physics, "potential" is defined as the work that is necessary to bring a unit mass from a given reference position to a particular point; consequently, it is the product of the force per unit mass and the distance. From the standpoint of soil moisture, if a definite mass of water could be pulled away from a given mass of soil, a certain amount of work would be required. This would be equivalent to the capillary potential times the mass of water, or the work to overcome the force of attraction of the water at that point. The capillary potential, therefore, is the work required to pull a unit mass of water away from a unit mass of soil. It is defined as the work required to move a unit mass of water, against capillary forces in a column of soil, from a free-water surface to a given point above this surface. Capillary pressures are negative because of the greater in-

ward attraction of the water molecules at the air-water interface. That is, the molecules within the liquid exert a greater attraction on the surface molecules than the air, causing a tension on the surface, which is a negative pressure. Consequently, the capillary potential is negative in sign.

Buckingham used metal cylinders, 48 in. long and $2\frac{1}{2}$ in. in diameter, which were filled with soil and brought in contact with a free-water surface, to investigate the relationship between the moisture content of the soil and the height above the water table. He developed the following theoretical analysis of the relation between the capillary potential, ψ , and the distance from the water table:

1. Assume an experimental setup where a column of soil of uniform packing is in a state of equilibrium as far as the movement of moisture is concerned. Evaporation is prevented at the surface and no further change in moisture distribution occurs within the column. Let x equal any given point above the free-water surface when the soil is saturated with water.

2. Move an infinitesimal mass of water, dm , from x to the point $x + dx$, where dx is an infinitesimal distance.

The work done by the applied force against capillary attraction in removing water from the soil at x is ψdm . ψdm represents the work that was released per mass of water dm when the soil was originally saturated at the point x above the free-water surface.

The work done by capillary attraction against the applied force, as water unites with the soil at the new level, $x + dx$, where the potential is equal to $\psi + \frac{\partial \psi}{\partial x} dx$, amounts to $\left(\psi + \frac{\partial \psi}{\partial x} dx \right) dm$. (In this expression, $\frac{\partial \psi}{\partial x} dx$ is a mathematical expression of the rate of change of the capillary potential with distance times the distance concerned, dx . Therefore, the total potential at this new point is equal to the original potential plus the amount it increases over the chosen infinitesimal distance.)

The total work done by the applied force against capillarity is given by the expression:

$$\begin{aligned} W_1 &= \psi dm - \left(\psi + \frac{\partial \psi}{\partial x} dx \right) dm \\ &= - \frac{\partial \psi}{\partial x} dx dm \end{aligned} \quad (27)$$

But, in moving the mass of water dm from x to $x + dx$, work had to be performed against gravity. Therefore,

$$W_2 = gdm dx \quad (28)$$

At equilibrium, however, $W_1 + W_2 = 0$. Therefore,

$$-\frac{\partial \psi}{\partial x} dx dm + gdm dx = 0$$

or

$$\frac{\partial \psi}{\partial x} = g \quad (29)$$

Not only does the capillary potential, ψ , depend upon the height above the free-water surface, but also the moisture content, Θ . Consequently, the change in ψ with respect to Θ is:

$$\begin{aligned} \frac{\partial \psi}{\partial \Theta} &= \frac{\partial \psi}{\partial \Theta} \cdot \frac{\partial x}{\partial x} \\ \frac{\partial \psi}{\partial \Theta} &= \frac{\partial \psi}{\partial x} \cdot \frac{\partial x}{\partial \Theta} \end{aligned} \quad (30)$$

Since $\frac{\partial \psi}{\partial x} = g$, then

$$\frac{\partial \psi}{\partial \Theta} = g \cdot \frac{\partial x}{\partial \Theta} \quad (31)$$

Equation 31 shows that the change of the capillary potential with moisture is equal to a constant, g , divided by the rate of change of moisture with height. Integration of equation 31 gives the formula:

$$\psi = gx + B, \quad (32)$$

where B is the integration constant. Since $\psi = 0$ when $x = 0$, then $B = 0$ and

$$\psi = gx \quad (33)$$

Equation 33 states that the capillary potential is proportional to the height above the water table. Consequently, if the variation in moisture content with height is known, the relation of capillary potential to moisture content is easily obtained.

Buckingham plotted $\psi/A = x$ against Θ and obtained a series of curves of the type shown in Figure 42. The curves have two major portions. At the lower end, the slope is small as the moisture content decreases rapidly with increasing distance above the free-water surface.

The curve then merges into a relatively steep portion where the slope

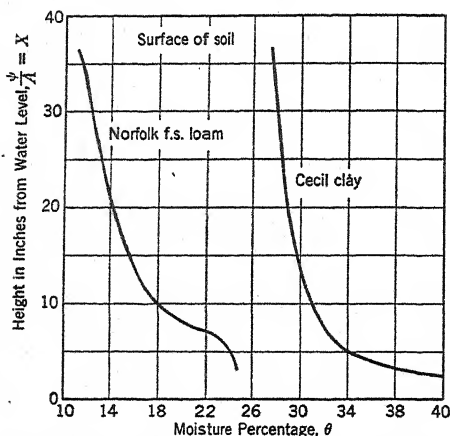


FIG. 42. Capillary potential-moisture curves of Buckingham (10).

is about the same for the different soils. These curves do not exhibit any distinct breaks. In other words, the force of attraction for water, or the work necessary to remove water from the soil (capillary potential), is a continuous function of soil moisture.

This concept of capillary potential can be more readily visualized if one compares it with the pressure potential or pressure-head differences and the flow of water through a level pipe. Such an analogy is illustrated in Figure 43. Let us consider the flow of water from reservoir, *R*, through the pipe, *ABCDE*. At the entrance to the pipe *A*,

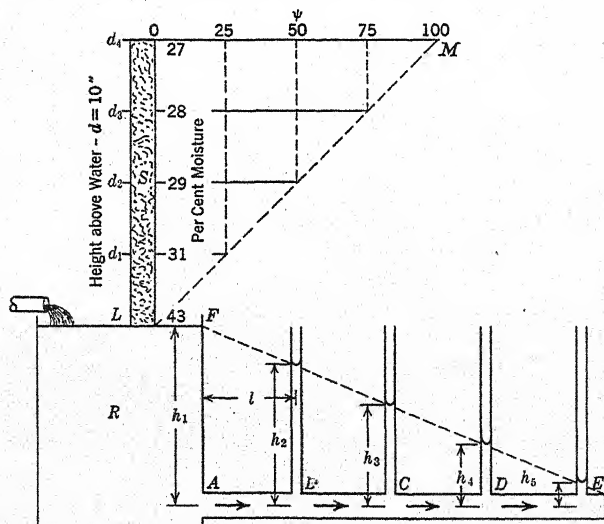


FIG. 43. Schematic comparison of the similarities between capillary and flow potentials.

the pressure head is equal to the weight of the water times the height, or wh_1 . At *B*, the pressure head is wh_2 . The difference in pressure

head between A and B is equal to $w(h_2 - h_1)$. The line FE shows the decrease in pressure head with distance from the reservoir. The driving force or pressure potential at A is equal to gh_1 ; at B it is gh_2 . The gradient in pressure potential between A and B is $g(h_2 - h_1)/l$. Since $(h_2 - h_1) = (h_3 - h_2) = (h_4 - h_3)$, this same potential gradient exists between equal lengths along the pipe. Therefore, the driving force between any two points is proportional to the potential difference.

Let us now analyze the capillary-potential relationships in the soil column, S , which is assumed to be that of the Cecil clay in Buckingham's experiments as illustrated in Figure 42. At the point L , where the soil is in contact with the free-water surface, the moisture content is 43 per cent and the capillary potential is zero ($\psi = 0$). At a distance d_1 (10 in.) above the surface, the moisture content is 31 per cent and $\psi = 25$; at d_2 , $\psi = 50$. The line LM shows the increase in capillary potential (increase in negative values since -50 is less than -25) with distance above the free-water surface. The capillary-potential gradient is $25/d$ (cm.) gm. per gm. Thus, it is seen that, whereas the pressure potential decreases positively with distance from the source of water, the capillary potential increases negatively (also a positive decrease) with distance. The two are analogous in that water movement is proportional to the potential difference between any two points.

Another simple analogy is to compare the capillary potential with the thermal or temperature potential in the flow of heat through a metal bar. Everyone is acquainted with the fact that heat flows from hot to cold areas. If a long metal bar is held between the fingers and touched against a hot object, heat will flow through the bar towards the cold end. If the temperature of the bar is measured at various points after the cold end has become slightly warm, there will be a gradual decrease in temperature from the hot to the cold end.

Suppose that the pipe AE in Figure 43 represents the metal bar, R is the hot object and the different tubes are thermometers placed in contact with the bar. Then, the line FE will correspond to the fall in temperature at the different points throughout the bar. The potential difference between A and B will be proportional to the difference in temperature between these two points. The same will be true between any other two points.

Let us return to the soil column, S , in Figure 43 and see what the various capillary potentials mean in terms of the energy with which water is held at the different moisture contents. At L the soil is

saturated, the capillary potential is zero, and practically no work is required to remove a gram of water from the soil. The value of ψ at d_1 is 25 gm. per gm. This signifies that a suction of 25 cm. of water is necessary to overcome the forces of capillary attraction and remove water from the soil at a moisture content of 31 per cent. At d_3 , with a moisture content of 28 per cent, a suction of 75 cm. of water is required to extract water from the soil. Thus, it is seen that the amount of water in the soil is a function of the energy with which the water is held. Moreover, the energy necessary to remove water is a continuous function of the moisture content, which makes any classification of soil water necessarily arbitrary. It is seen from the foregoing discussion that the investigations of Buckingham have pointed the way for a more fundamental concept of soil-moisture relationships.

The work of Buckingham was not given serious consideration until his original ideas were expanded by Gardner (18, 19) in 1920, who suggested that the capillary potential gives a new interpretation of the various soil-moisture constants employed by Briggs and others. He pointed out that some experimental evidence indicates that the capillary potential is a linear function of the reciprocal of the moisture content, at least over a considerable range of the curve, according to the relationship:

$$\psi = \frac{e}{\theta} + b$$

where e and b are constants. This means that the potential-moisture curve is hyperbolic in nature. The curves of Richards in Figure 44 bear out such a possibility. Commenting on these curves, Gardner (20) says, "Their simple form suggests that there would be required but a small number of parameters for their representation. It seems likely, also, that the fundamental physical significance of these parameters would make them suitable for identification of a soil and, therefore, for a basis of soil classification." In other words, he is hopeful of expressing the capillary-potential-moisture relationship in terms of definite soil physical properties.

Richards (41) has expanded the concepts of Gardner, especially from the point of view of the technique of measurement. In order to hasten the attainment of equilibrium during potential determinations he has used a $\frac{1}{2}$ -inch layer of soil on a porous plate which is sealed onto the top of a rectangular reservoir. Various tensions are possible by changing the amount of suction on the water in the reservoir. When the moisture content of the soil has reached equilibrium, the soil has a

potential at that point equivalent to the suction applied. Several typical curves of Richards are illustrated in Figure 44. These data show that a coarse-textured soil exhibits a high negative potential at a low moisture content. Finer-textured soils contain much more water at the same potential. This is due to the larger number of contacts in the finer-textured soils, thereby reducing the amount of moisture at each of the contact points with a corresponding decrease in the radius of curvature of the water menisci in the pores.

If these soils were placed in tubes in contact with a free-water surface and permitted to come to equilibrium, the sand soil would have

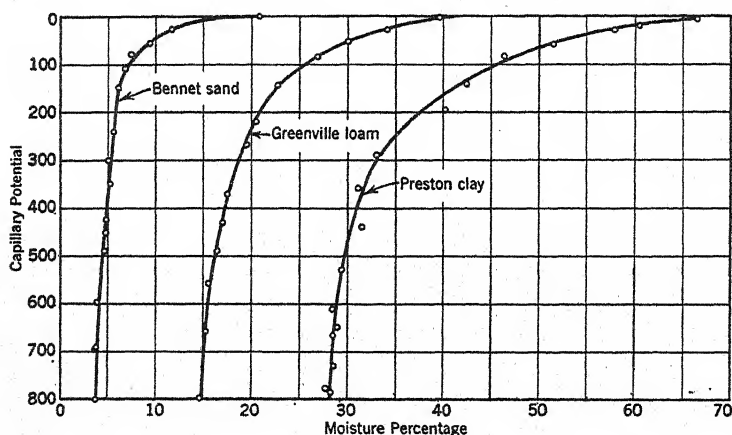


FIG. 44. Tension-moisture curves of Richards (41).

a moisture content of about 4 per cent at 800 cm.; the clay soil would contain about 28 per cent. The curves also indicate a tendency for the moisture content to become constant at the more negative values of the capillary potential. This effect has been observed by Lebedeff (31) in studying the distribution of water in columns of sand of various heights as a function of height, after they had been saturated and allowed to drain at the bottom. The moisture content decreased with height up to about 40 cm., above which it remained practically constant. Lebedeff has called the water content of this layer the "maximum molecular moisture-holding capacity" of the sand. Forces of molecular cohesion as influenced by the properties of the surface are considered to be the cause of water retention at this point.

Richards has pointed out that moisture content, size of particle and state of packing affect the value of the capillary potential through

their effects upon the curvature of the moisture films. The reader is referred to his excellent paper for a rather complete discussion of the capillary potential and its usefulness to soils investigations.

Veihmeyer and Edlefsen (56), in reviewing the field of energy relationships associated with soil-moisture problems, emphasize the fact that the energy necessary to remove water from the soil is a continuous function of soil moisture. They also suggest that the accuracy of measurements now available justifies the assumption of a linear relation for the major parts of the potential-moisture curves, especially between the moisture contents of the field capacity and permanent wilting. Although this is not strictly true in the mathematical sense, there may be some practical justification for such an assumption, at least on the basis of many experimental observations.

Wadsworth (56), in reply to the above suggestions, hints that the energy-moisture curve may be nothing more than the "composite of several curves in the same coordinates." This possibility should not be overlooked in an interpretation of the energy-moisture curves. Even though the energy that must be expended to remove water from the soil from saturation to dryness is apparently a continuous function of the moisture content, it does not necessarily follow that there is no change in the nature of the factors responsible for the attraction and retention of water somewhere along the curve. This particular phase of soil-moisture investigations needs to be studied further before it can be safely said that the capillary potential is a single-valued function of soil moisture. Wheeting (56) has offered the interesting possibility that there are at least two zones in which the factors responsible for the energy of attraction are different. The dry zone is characterized by a domination of molecular forces at the solid-liquid interface at the surface of the particles. The wet zone is featured by tension forces operating at the air-liquid interface within the soil pores.

These viewpoints call attention to the fact that additional analysis of energy-moisture relationships in soils is necessary before any wide correlation between the capillary potential and soil physical properties can be achieved. The introduction and use of the energy concept of soil-moisture retention and movement, however, have placed the problem of soil-water relationships on a fundamental basis.

MEASURING THE CAPILLARY POTENTIAL. Various methods may be used to obtain data from which the potential can be calculated. The more important techniques that may be employed are: vapor pressure, freezing-point depression, dilatometer, moisture absorption by seeds, water distribution in long columns and water distribution under

centrifugal force or under tension in conjunction with porous clay cells.

Vapor-pressure measurements have been discussed in Chapter III. Data from vapor-pressure determinations permit the calculation of the capillary or moisture potential within the dry range on the moisture scale; that is, throughout the range where a tension of about 10 atmospheres or over is necessary to remove water from the soil.

Freezing-point depression and dilatometer (6) data can be used to determine the potential throughout the moisture range that is the most favorable for plant growth, namely, from about the moisture content at permanent wilting up to that of the moisture equivalent. Schofield and DaCosta (47) have modified the usual freezing-point depression technique to make possible calculations of capillary tension over this moisture range.

Shull (52) has used *Xanthium* seeds, placed in contact with soils containing known amounts of water, to measure the force with which water is held in soils. The seeds are essentially natural osmometers. Values obtained by him indicate that this technique gives fair results at low moisture contents.

The use of long soil columns for measuring the capillary potential has been discussed in conjunction with the investigations of Buckingham. The major difficulties with this approach are the length of time that is required to attain equilibrium and the necessity of having columns sufficiently long to obtain values at tensions within the moisture range that is optimum for plant growth. This technique provides data for the extreme wet end of the moisture scale. Experimental difficulties, however, have prevented any wide usage of this type of measurement.

The centrifuge offers many possibilities for measuring the retention of water against definite centrifugal forces. This fact has been emphasized by the extensive use of the moisture-equivalent centrifuge to determine the amount of water that is held in the soil against a force 1000 times that of gravity. Lebedeff (32) has shown that the moisture content of all soils decreases gradually with increases in the centrifugal force from 400 to 18,000 times gravity. With a centrifugal force between 18,000 and 70,000 times gravity, the moisture content remains almost constant. The curve showing moisture content as a function of centrifugal force is almost linear at these higher forces. Lebedeff has assumed that the amount of water retained against a force 18,000 times that of gravity represents that water which is held because of the molecular forces of cohesion in the surface of the soil particles. Olmstead (39) has been able to centrifuge soils in an air-driven machine at

a speed equivalent to a force 300,000 times that of gravity. The amount of water retained against such a force is nearly the same as that of the hygroscopic coefficient. These results point out that the force of centrifugation may be used to measure the moisture potential. Recently, Russell and Richards (45) have perfected a special technique whereby values of the tension between the moisture equivalent and the wilting point can be determined with the centrifuge.

The most widely known method for measuring the capillary or moisture potential is based upon the so-called suction force of the soil for water. The essential features of the technique are the placing in the soil of a porous-clay cell or cup which is sealed onto a mercury manometer, filling the cell and part of the manometer with water and allowing the system to come into equilibrium with the soil. As the soil

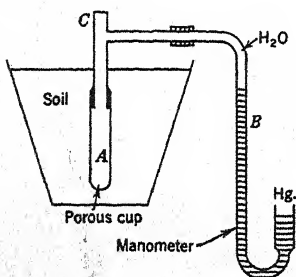


FIG. 45. A simple tensiometer for determining the tension-moisture curve (Richards, 41).

dries out, water leaves the cell, enters the soil and the mercury rises in the manometer. As the soil becomes wet, and the tension of the water in the soil is less than that in the cup, water enters the cup again which causes the mercury in the manometer to fall. When the moisture in the soil is in equilibrium with the water in the cup the potential of the soil at that given moisture content is equal to that represented by the reading of the manometer. Gardner and his associates first referred to this type of an instrument as a "capillary potentiometer." Later,

Richards and Gardner (42) suggested the term "tensiometer" for the sake of brevity and to describe more or less the function of the apparatus. A simple type of tensiometer is illustrated in Figure 45. The porous cup, *A*, is sealed onto a manometer tube, *B*; the cup and part of the manometer are filled with water through the opening, *C*, which is tightly closed with a stopper; water is in direct contact with the mercury. The porous-clay cup is then placed into the soil in such a way as to insure close contact between cup and soil. The movement of water into or out of the cup, as determined by the difference in tension between the water in the cup and in the soil, is measured by the change in height of the mercury column in the manometer.

Richards and Gardner have described three general types of tensiometers. These are the indicating, recording and differential types. The first is similar in principle to the one illustrated in Figure 45; a vacuum gauge may be substituted for the mercury manometer. The

second is the same as the indicating type except that a self-recording vacuum gauge is used to obtain a continuous record of the tension. Differential tensiometers consist of two porous cups that are filled with water and connected to two sides of a mercury manometer. The difference in tension at the two cups is thereby measured; this makes possible the detection of the direction of movement of water in the soil. The reader is referred to the work of Richards (43, 44) for a detailed discussion of various types of tensiometers.

The tensiometer provides data on the wet end of the energy-moisture curve. The maximum tension cannot exceed one atmosphere. When the potential of the soil water exceeds this value, air enters the cup and the instrument is no longer operative. There is a considerable hysteresis effect. A soil that is drying out has a different moisture content at the same reading of the manometer than the same soil when it is being wetted. Nevertheless, this technique offers many possibilities for studying soil-moisture changes from values about the moisture equivalent to capillary saturation.

PRESSURE DEFICIENCIES IN RELATION TO THE NATURE OF THE PORES. Almost concurrently with the investigations of Gardner and his associates in America, Haines (21) of Rothamsted developed the concept of moisture retention and movement in an ideal soil on the basis of the suction force necessary to move an air-water interface throughout the cellular-like pores of the soil. Haines based his concept on the classical investigations of Slichter (53) and King (29) on the nature of the pore space in an ideal soil in open- and close-packed arrangement. Slichter calculated that an arrangement of spheres in an open-packed manner would have a maximum porosity of 47.64 per cent. In this type of packing (Figure 27*a*), each sphere touches six others and the lines joining the centers of these spheres form a cube with a side equivalent to $2r$ or twice the radius of a sphere.

In the close-packed system (Figure 27*b*), with twelve points of contact per sphere, the unit cell is a rhombohedron formed by passing planes through the centers of eight contiguous spheres. The length of the side is twice the radius of the sphere ($2r$) and the face angles are 60° and 120° . Such a unit is shown in Figure 46*a*; a plaster-cast reproduction of the pore space between the spheres is illustrated in Figure 46*b*. It is noted that the pore space is bounded by curved surfaces and that the so-called capillary tubes are triangular in cross section. Moreover, there are maxima and minima cross-sectional areas throughout the pore volume. In other words, the larger cells are connected by more constricted portions of the soil-pore space.

Haines first considered a single water wedge between two spheres

in contact, in order to develop the concept of the pressure deficiency under the curved water surface. (See Figure 27c.) The development of this relationship has been given in conjunction with the discussion on cohesion. (See page 106.) He employed the nomenclature of

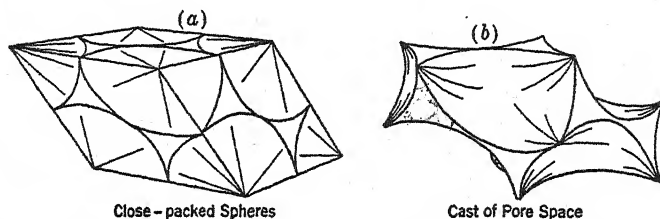


FIG. 46. The nature of the pore space in an assemblage of close-packed spheres (after Slichter, 53).

Versluys (60) to characterize the different stages in the filling of the soil pores with water. These stages are:

1. THE PENDULAR STAGE—water is present as rings around the points of contacts between spheres. The edges of the rings come in contact as the moisture content increases and the rings grow larger.
2. THE FUNICULAR STAGE—water wedges are in contact and form a continuous network throughout the system. Likewise, the air throughout the pores is continuous. As the moisture content increases the pores begin to close at the narrowest parts, enclosing air bubbles within the cells.
3. THE CAPILLARY STAGE—all cells are saturated with water and the pressure deficiency is zero.

Haines used sand and uniform glass spheres on a sealed filter paper in a Büchner funnel, which was connected to a burette-manometer, to measure the moisture content as a function of pressure deficiency. The apparatus was filled with water and the desired pressure deficiencies, or suction values, were obtained by adjusting the height of the burette. Typical results with this technique are illustrated in Figure 47. It is noted that the drying and wetting curves differ considerably; there is a marked hysteresis. The various portions of these curves have been explained as follows: Consider first the drying curve, *OABCDE*. At *A*, the entry of air is first observed and is rather general at *B*. From *B* to *C*, water is being removed from cells with small entrances; there is some decrease in funicular water. All the funicular water is removed at *E* and the pendular stage of water

is dominant. It is evident that there is an overlapping of these different stages and that no distinct demarcation of the various forms of water is possible.

Consider next the wetting curve *DFGHK*. The pendular stage is represented by the reversible portion *FD*. At *F*, enough water has entered the soil to close part of the smaller cells; the closing of average-sized cells takes place at *G*; saturation is complete at *H*. It is seen that at the highest degree of saturation, *HK*, about 10 per cent of the pore space is filled with entrapped air. The differences between the

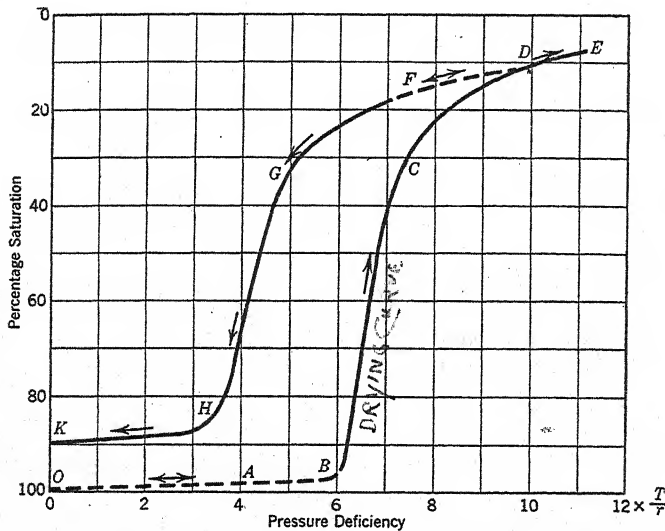


FIG. 47. Pressure deficiency curves of Haines (21).

drying and wetting curves, therefore, are associated with variations in the pressure deficiencies in a cell filling with water as compared with one that is draining. These hysteresis effects present some of the greatest difficulties in using pressure-deficiency or capillary-potential techniques for measuring soil moisture in the field.

✓ THE pF OF SOIL WATER. Schofield (46) has simplified the concept of energy relationships in soil-water problems by introducing the idea of using the logarithm of the capillary potential, pF , to express the energy with which water is held by soils. The use of pF has several advantages over the employment of capillary potential as such. In the first place, it eliminates the word "capillary" from the terminology of energy relationships. This is highly desirable since surface-tension forces are only a part of the total forces involved in

the attraction of water by soils. The growing use of moisture potential among soils investigators is evidence of the need of such a change in the definition of terms. Second, by analogy with pH , the symbol p indicates a logarithm; the character F is supposed to suggest free energy. Therefore, by defining pF as the logarithm of the height in centimeters of a liquid (water) column that is necessary to produce the desired suction, it is possible to transfer the scale to any liquid. Third, the use of the logarithmic scale permits the designation of the relation of tension to moisture content on one graph. In this way, it is possible to show the relationship of the various so-called soil-moisture constants to each other. This possibility represents the real contribution of Schofield in introducing the pF concept.

Schofield has used data of other investigators, in addition to his own experimental results, for calculating the pF curve. For example, the typical curves illustrated in Figure 48 on the Greenville loam from Utah represent vapor pressure and centrifuge data of Thomas (54), as well as results by Richards (41) for wetting against suction on a porous plate. The pF can be calculated from vapor-pressure and freezing-point determinations by the following formulae:

$$pF = 6.5 + \log_{10} (2 - \log_{10} h)$$

$$pF = 4.1 + \log_{10} t$$

where h is the relative humidity in per cent at 20° C. and t is the freezing-point depression in degrees centigrade.

These curves show that the various soil-moisture constants that are so commonly used to characterize moisture relationships are on the same curve. The moisture equivalent occurs at a pF value of about 3, although Russell and Richards (45) have recently indicated that it is nearer 2.7; the moisture content at field capacity corresponds to a pF of about 3.2; permanent wilting takes place at a pF of 4.2, or at 10 times the tension of the moisture at field capacity. Thus, it is possible to draw an approximate pF curve from the hygroscopicity at 50 and 99 per cent relative humidity, the moisture equivalent, and the moisture content at capillary saturation at a tension of 2 to 5 cm. Dickinson fine sand represents a complete moisture-tension curve. The curves also point out that at the same tension a soil is more moist if it is being dried than if it is being wetted.

By way of summary of the discussion on the concepts of soil moisture, it should be pointed out that the essential feature in the change from the capillary-tube hypothesis to the energy point of view

has been the shifting of emphasis from a more or less static to a dynamic explanation of soil-water relationships. The earlier stages of soil-moisture research were more concerned with apparent distribu-

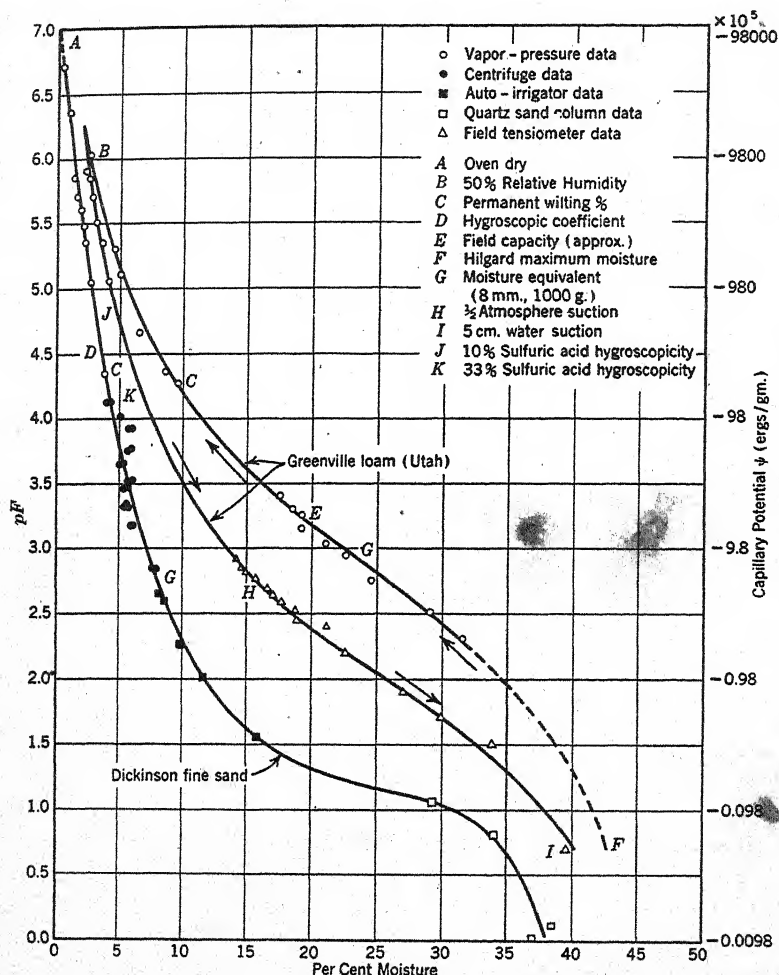


FIG. 48. The pF of soil water (Schofield, 46, and Russell and Richards). The curve for the Dickinson fine sand is through the courtesy of Russell and Richards (data as yet unpublished).

tion of water in the soil; recent stages place the major emphasis on the distribution of the forces responsible for the retention and movement of water.

THE MOVEMENT OF WATER

Movement of Water in the Liquid Phase. The movement of soil water through a given volume of soil must take place through the soil pore space. Movement of water through these pores is brought about by the action of gravity or capillary pull, either alone or in combination. According to the dominance of the moving force, the type of water movement may be discussed from two points of view: (1) that which moves in the larger pores primarily through the action of gravity, or movement in a saturated soil, and (2) that which moves primarily through the action of capillary forces from surface to surface or in the small pores in the presence of numerous air-water interfaces, or movement in an unsaturated soil.

MOVEMENT IN UNSATURATED SOILS. An unsaturated soil is one in which the larger pores are filled with air and, consequently, where movement is closely dependent upon a large number of air-water interfaces. The distribution and movement of water may be laterally, vertically upward, vertically downward or at any angle between the vertical and horizontal. Soil-moisture movement under these conditions will be discussed from the viewpoint of the older concepts of capillarity as well as from the more recent analogies to the flow of heat or electricity.

Capillarity. The fundamental basis of the capillary-tube hypothesis, with respect to the movement of capillary water in soils, is the well-known height of capillary-rise equation:

$$h = \frac{2T}{gDr} \quad (34)$$

where h is the height of the meniscus above the water level, T is the surface tension and D the density of the liquid, g is the acceleration due to gravity and r is the radius of the capillary tube. Interpreting this equation in terms of the soil, we see that the height of rise is inversely proportional to the radius of the pores. The presence of many small pores in the soil would therefore suggest considerable capillary movement of water in soils. Keen (26) has calculated the following theoretical heights of rise of water in ideal soils: fine gravel, $\frac{1}{3}$ foot; coarse sand, $1\frac{1}{2}$ feet; fine sand, $7\frac{1}{2}$ feet; silt, $31\frac{1}{4}$ feet; fine silt (0.01–0.002 mm.), 150 feet; and clay, over 150 feet. These calculations presuppose that sufficient time is permitted for equilibrium. The attainment of such large distances of movement, either experimentally in the laboratory or under natural conditions, is almost impossible.

Some investigators have claimed that water has risen long distances by capillarity from deep-water tables but there is little experimental evidence to justify such claims.

The movement of water by capillary action has interested soils investigators for a long time. In 1884, Wollny (63) studied various factors that affected the capillary movement of water in soils. He found that the rate of capillary rise in a loam soil increased with temperature, with the looseness of packing and with the original moisture content of the soil. He also observed that the rise was faster in a column of sand particles of mixed sizes than in columns having par-

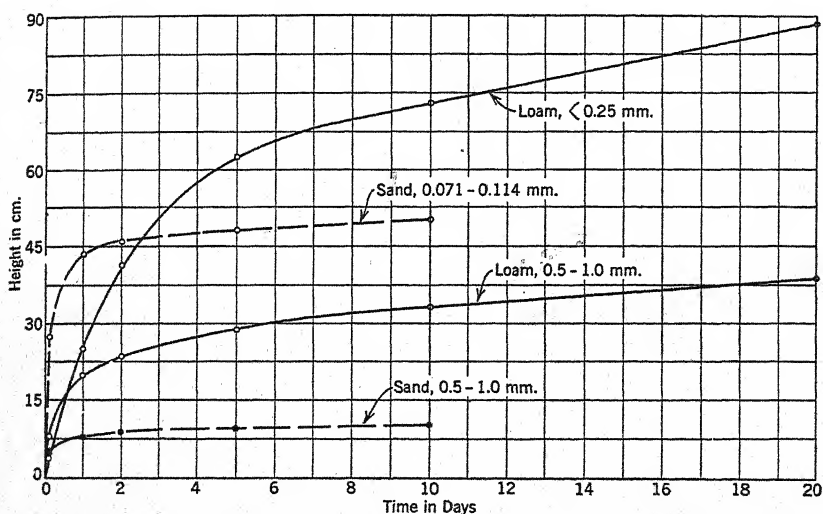


FIG. 49. Rate of capillary-rise curves plotted from the data of Wollny (63).

ticles of uniform size. He came to the conclusion that soil capillaries from 0.05 to 0.10 mm. in diameter conducted water the most rapidly and that capillary movement reached a minimum limit with quartz particles larger than 2.0 mm. Several interesting curves from Wollny's data are illustrated in Figure 49. They show that coarse-textured soils have a rapid initial rate of capillary movement but that the finer-textured one eventually has the highest rise. This is in conformity with most measurements of this type. The well-known experiments of Loughridge (33) in California, which were carried out only 7 to 8 years after those of Wollny, point out the same tendencies. These curves also indicate the difference between the capillary movement in sand grains and loam granules of the same size. The latter permit greater movement of water through the pores between the granules.

Translated into practical terms, this means that a granular soil possesses capillary properties (water-holding capacity) within the granules and at the same time permits the ready movement of water in the larger pores between them.

King (30) showed that capillary movement through a wet soil is faster than through one that is dry. He emphasized this fact by advocating that a thin layer of dry soil should be maintained on the surface to act as a mulch for preventing evaporation.

Harris and Turpin (22) presented some interesting data on the capillary movement of water. They showed that the greatest rise and descent of water into a dry soil from a moist one always took place in the case of the greatest initial moisture content of the source. Capillary movement downward was slightly faster than upward or laterally. The rate of movement from one soil type into another varied considerably with differences in texture. Only a small rise was observed from a moist clay soil into a dry loam that was brought in contact with the clay; an exceedingly large and rapid rise occurred from moist sand into a dry loam. Although Harris and Turpin did not interpret their data in terms of modern ideas on capillary movement of water, nevertheless, the results they obtained, by rather clever techniques, can readily be explained on the basis of present concepts.

Practically all experimental results on capillary rise point towards the fact emphasized by Johnson in 1878, that the "capillary properties of the soil must be evaluated in terms of rate as well as distance of movement." The capillary-tube hypothesis has emphasized distance of movement rather than the rate. Experiments show, however, that the rate of capillary rise is very slow in those soils where the pore size mathematically suggests a high rise. In other words, these older concepts have indicated that there is a capacity factor and a conductivity factor in capillary phenomena; they seldom have been recognized as such in the analysis of the experimental data.

Analogy of Capillary Movement with Heat Flow. Buckingham, in his original work on the capillary potential, defined the capillary flow of water as follows: "We must think of it as a current of water through the soil; and as a measure of the strength of the capillary current at any point we may take the amount of water which passes in one unit of time through a unit area of an imaginary plane surface perpendicular to the direction of motion. We may measure the current strength or current density in centigrams per day per square centimeter." This definition finds its counterpart in explanations of the flow of water in tubes, the flow of heat and the flow of electricity.

The rate of flow of water through a pipe is proportional to the

difference in pressure head times the conductance. If we were interested in determining the rate of flow across any given cross-sectional area in the pipe illustrated in Figure 43, we should say that this rate would be proportional to the conductance times the potential gradient, that is, the fall in pressure head per unit length times g .

The use of the term "potential gradient" necessitates that it be clearly distinguished from "potential." The latter has magnitude only and is not concerned with direction. The former, however, has both magnitude and direction; that is, gradient implies that there is a change in one direction or another. *The direction of the potential gradient in which the potential increases is considered positive.* In other words, the potential gradient is the change in potential per unit distance in the direction of the maximum rate of increase of potential.

Since flow always takes place from a higher to a lower potential, that is, in opposite direction to the potential gradient, the rate of flow of water through a pipe is expressed as follows:

$$\text{Rate of flow} = -K \text{ Grad } \pi \quad (35)$$

where K refers to the conductance of the pipe and $\text{Grad } \pi$ is the potential gradient. The conductance of the pipe varies directly with the fourth power of the diameter and inversely with length (d^4/l).

According to Fourier's law for the flow of heat:

$$q = -C \text{ Grad } \theta \quad (36)$$

where q is the quantity of heat crossing unit cross-sectional area perpendicular to the direction of flow in unit time, C is the thermal conductivity and $\text{Grad } \theta$ is the thermal potential gradient or the change in temperature per unit distance in the direction of the greatest change in temperature.

In turning to the field of electricity, we find that according to Ohm's law:

$$i = -C_1 \text{ Grad } V \quad (37)$$

where i is the current density, C_1 is the specific conductivity and $\text{Grad } V$ is the change in electrical potential per unit distance.

If we can analyze the capillary flow of water according to this type of reasoning, taking into consideration the meaning of the capillary potential as discussed on page 197, we obtain the following expression:

$$Q = -K \text{ Grad } \psi \quad (38)$$

where Q is the capillary-current density or the mass of water which passes in one second through one square centimeter of an imaginary

plane perpendicular to the direction of flow, K is the capillary conductivity and $\text{Grad } \psi$ is the capillary-potential gradient, that is, the amount by which ψ increases negatively per centimeter in the direction of flow, because of the fact that the moisture content of the soil decreases in that direction.

Equation 38 expresses the movement of water under the driving force of the capillary potential only, as horizontal flow, for example. When movement takes place upward or downward, the force of gravity enters into the picture. The flow of water in a soil may be expressed, according to the Darcy equation:

$$v = -K \text{ Grad } \Phi \quad (39)$$

where $\text{Grad } \Phi$ represents the change in the total water-moving force per unit distance and K is the specific conductivity or the amount of water which will flow in one second across a unit cross-sectional area of soil perpendicular to the direction of flow, when the value of Φ changes at the rate of one unit per centimeter. In this case, $\text{Grad } \Phi = \text{Grad } \psi + \text{Grad } \phi$, where $\text{Grad } \phi$ is the gravitational potential gradient. Since $\phi = gh$, $\text{Grad } \phi$ is equal to g .

The capillary conductivity depends upon the kind of soil, its state of packing and the moisture content. Those particular soil properties that affect capillary movement are included in the evaluation of K . Buckingham showed that the capillary conductivity increased with the moisture content and decreased with the size of the soil pore. The extremely low conductivity at small moisture contents has been observed and emphasized by numerous investigators. Moore (38) has shown that the capillary conductivity, or permeability according to his terminology, is a maximum at or near saturation. It then decreases rapidly with the moisture content to about the moisture equivalent, where the permeability is very low and remains practically constant as the moisture content is further decreased. At this moisture content, the rate of change of conductivity with moisture is approximately zero and is interpreted as representing that moisture content at which the water films in the soil become discontinuous. Capillary permeability, therefore, is approximately zero and any further moisture distribution must take place through movement in the vapor phase.

The texture and structure of the soil affect capillary conductivity as they influence the number, size and continuity of the pores. Moore (38) has measured the capillary conductivity of soils with different textures and has found that at a pF of 2, for unsaturated flow, the various soils are arranged in the following order with regard to capillary

permeability: Sand < fine sandy loam < light clay < clay. For saturated flow the order of permeability is reversed. The moisture content at which capillary permeability becomes zero for each of these soils is: sand, 5.0 per cent (pF 1.90); fine sandy loam, 20.4 per cent (pF 1.96); light clay, 24.6 per cent (pF 2.08); and clay, 25.9 per cent (pF 2.15). These results point out that the water films in sands become discontinuous at much lower tensions or moisture contents than in clays. This is due to the fact that clays possess a large number of small pores in contrast with a smaller number of larger pores in sands.

Gardner (18) has used the term "capillary-transmission constant" to express the capillary conductivity of soils. He has considered it to be related to the effective radius of the particles and the porosity of the soil. The effect of packing on conductivity is shown in the following data obtained by Gardner with the Greenville soil:

TYPE OF PACKING	CAPILLARY CONDUCTIVITY
Extremely loosely packed.....	-1.8×10^{-3}
Well packed by previous wetting and drying....	-5.4×10^{-3}
Packed dry by systematic tapping of tube.....	-7.4×10^{-3}
Field soil in natural structure.....	-8.7×10^{-3}
Mean.....	-5.8×10^{-3}

The conductivity in the field soil includes both gravitational and capillary flow. These data show that an optimum porosity exists since the loosely and closely packed samples exhibit lowest conductivity. In the former, the pores are apparently too large to permit establishment of a continuous moisture film. In the latter, the pores are so small that there is considerable resistance to flow.

From the above discussion it is noted that the capillary flow of water in soils is determined by the driving force or potential gradient and the conductivity. Equation 39 indicates that a rapid rate of flow takes place only when both the conductivity and potential gradient are large. A slow movement of water results if either or both of these factors are small. The greater the difference in moisture content between two cross-sectional areas in a soil column, the greater will be the potential difference (driving force) and the more rapid will be the movement of capillary water. This fact is well illustrated by the curves of Gardner and Widtsoe in Figure 50, which were plotted from data by Harris and Turpin (22). These curves are self-explanatory and show that the more moist the reference soil, the more rapid is the movement of water into the dry soil with which it is in contact.

Moore (38), in studying the advance of the wetting front as water

risers in a soil column from shallow-water tables, has characterized the wetting front as follows:

1. Water advances in a front from wet to drier soil under the influence of capillarity. Beyond the front, the soil remains apparently dry, and immediately at and behind the front the soil is apparently completely wetted. Macroscopically, there is a sharp line of demarcation between the obviously wet soil and the obviously dry soil.

2. The moisture content of the wetting front determined by sampling is constant, indicating a constant potential and a constant radius of curvature.

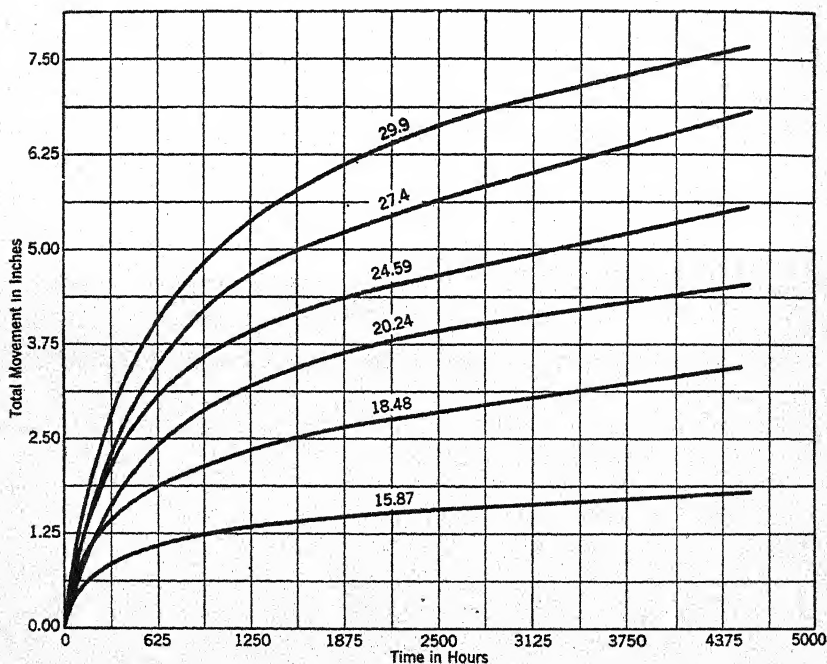


FIG. 50. Rate of movement of water into a dry soil from moist soils containing the percentage of water shown (after Gardner and Widtsoe, *Soil Sci.*, 11 : 230, 1921).

These experimental observations indicate that beyond the wetting front water films are either discontinuous or not present and that they are continuous behind the front. This means that the capillary conductivity of a soil is practically zero at moisture contents lower than that of the wetting front, since there is no continuity of moisture films. As Moore has pointed out, these facts show that the magnitude

of the potential gradient from a dry soil to the wetting front has little or no influence on the rate of movement of this front. It is true that a wet soil requires less water than a dry soil to establish continuous water films. The wetting front is advanced by flow through continuous films back of the front and is proportional to the potential gradient within this area.

✓ *Types of Capillary-water Movement.* Richards (41) has called attention to three common types of capillary movement of water. These are (1) the movement of precipitation or irrigation water downward through a comparatively dry soil, (2) movement of water upward from a saturated level or water table and (3) the movement of water in the horizontal direction.

The movement of capillary water downward takes place under the combined influences of the gravitational-potential gradient (Grad ϕ) and the capillary-potential gradient (Grad ψ). If evaporation is prevented at the surface, such downward movement will continue until the soil is drained or until equilibrium is attained with an impermeable layer or a saturated water table. Alway and McDole (1) found that water moves downward in a soil column, that is wetted from above and covered to prevent evaporation, until the moisture content behind the wetting front is reduced to about 2 to 3 times the value of the hygroscopic coefficient (99 per cent relative humidity). On the basis of the pF curves in Figure 48, their results suggested that equilibrium is obtained at a pF of about 3.0. This roughly corresponds to the moisture content where Moore considers capillary permeability to equal zero. It also conforms to the "field capacity," at term suggested by Veihmeyer and Hendrickson (58) to designate that moisture content of the soil after downward movement has "materially decreased." ✓ The interesting feature of the investigations of Alway and McDole is that each soil retained the same amount of water irrespective of its location in the soil column; six widely different soils were used.

Upward movement of capillary water takes place against the gravitational potential when the capillary-potential gradient is of sufficient magnitude to cause movement in this direction. Upward flow from a water table is maintained in the presence of evaporation. There has been a considerable difference in opinion as to the distance upward that water moves by capillarity as the surface layers become drier than the subsoil. McGee (36) estimated that, in the Great Plains, 6 inches of water annually are brought to the surface by capillarity from a depth of 10 feet. On the other hand, Alway and McDole of Nebraska came to the conclusion that the amount of water which the deep subsoil contributes to the growth of annual plants

is of no practical importance. They found that the movement of water upward from below 12 inches was exceedingly slow after the moisture content was reduced to about the field capacity. It should be emphasized again that these observations confirm the more recent findings of Moore on the relation of the potential to capillary permeability. Rotmistrov of Russia stated that water that penetrates beyond 16 to 20 inches does not return to the surface except by the way of plant roots.

Keen (28) has demonstrated that upward movement of water by evaporation lowered the water table in a coarse sand 35 cm. (about 14 inches) and in a heavy loam soil 85 cm. (about 34 inches) during 6 to 7 months. After this time, the rate of lowering of the water table was exceedingly slow. He concluded that the soil possessed only a limited capacity to supply moisture to the plants at depths lower than 3 feet below the root zone. Recent data reported by Richards and his associates (43, 44) suggest an upward water-moving field in the Marshall silt loam to a depth of about 24 inches. They obtained these results with tensiometers that were placed at various depths in a fallow plot.

There are, therefore, sufficient data to show that the upward rise of water by capillary action under conditions permitting evaporation is not as significant as the older points of view led one to believe. Emphasis is now being placed upon the limitations of capillary flow in explaining losses of water at the surface of soils by evaporation and the availability of water to plants.

Lateral or horizontal movement of water takes place under the influence of the capillary-potential gradient only. Harris and Turpin (22) have shown that lateral capillary flow is slightly less than downward movement. In most instances, the practical importance of lateral movement is definitely limited. An interesting example of differences in the horizontal flow may be found in the variations in the moisture content of the various horizons of the soil profile in a roadcut. As the soil dries out in the spring, or after a prolonged wet period, the heavier subsurface layers always appear dry much sooner than the immediate surface or the deeper horizons. This may be explained by the higher tension and slower conductance of the water in the clay layer. Evaporation at the immediate surface is much faster than the lateral movement of moisture.

MOVEMENT IN SATURATED SOILS. In saturated soils, water movement takes place throughout a soil-pore space that contains little, if any, air. In most instances, there is probably some entrapped air within the pores. It is difficult to remove all the air from a soil

column, even under carefully controlled laboratory conditions. In this discussion, however, we will assume that flow takes place through a completely saturated soil.

According to Darcy's law (1856) the velocity of flow of water through a column of soil is directly proportional to the difference in pressure head and inversely proportional to the length of the column:

$$v = k \frac{h}{l} \quad (40)$$

where v is the velocity in cubic centimeters per second, h is the difference in pressure head in centimeters, l is the length of the column in centimeters and k is the proportionality constant.

In order to use this equation for columns of various cross-sectional areas, it can be multiplied by the factor, F , which is the area in square centimeters. When h/l is unity and the area is one square centimeter, $v = k$ and k is then the coefficient of permeability or the permeability constant.

Slichter (53), in studying the movement of ground water in soils, developed the following formula to express the quantity of water per second that is transmitted through a soil column:

$$q = \frac{10.22pd^2s}{\mu hk} \quad (41)$$

where q is the quantity transmitted in cubic centimeters per second, p is the difference in pressure head in centimeters, d is the mean diameter of the soil grains in centimeters, s is the cross-sectional area in square centimeters, h is the height of the soil column in centimeters, μ is the coefficient of viscosity of the liquid in poises, k is a constant.

It is noted that the essential difference between equations 40 and 41 is the attempt to introduce a variable soil characteristic into the formula by Slichter. This is due to Slichter's ideas concerning the nature of the soil-pore space.

The permeability or transmission constant is a function of soil properties, particularly the size of the particles and the amount and nature of the soil-pore space. Slichter defined the transmission constant as the quantity of water transmitted in unit time through a cylinder of soil of unit length and unit cross section under unit difference in head at the ends. The value of the constant was given as $15.3d^2/k$. Slichter emphasized the fact that all formulae for flow are inadequate that do not take porosity into consideration.

Zunker (65) has recognized the importance of soil properties to

permeability and has used the relationship in equation 42 to calculate the transmission rates of soils.

$$Q = \frac{\mu h}{l U^2 \eta} \cdot \left(\frac{p_0}{1 - p} \right)^2 \cdot F \quad (42)$$

where Q refers to the amount of transmitted water, h the pressure difference, η the coefficient of viscosity, l the length of the column, U the effective specific surface, μ the type and arrangement of the particles, p the total pore space, p_0 the tension-free pore space and F the cross-sectional area. The value of μ varies from 2.3 for round particles to 0.5 for disc-shaped particles. Inasmuch as most soils contain particles of various shapes, the value of μ is usually taken as 1.0.

The tension-free pore space is calculated according to the formula:

$$p_0 = p - \frac{w}{100} (1 - p)s \quad (43)$$

where w is the inactive pore space occurring in 100 gm. of dry soil and s is the specific gravity of the soil. The value of w is taken as twice that of the hygroscopicity. In terms of the energy with which water is held in the soil, this value would be equivalent approximately to the moisture content at the field-capacity.

On the basis of this concept, Zunker has interpreted the Slichter transmission constant to vary directly with the square of the tension-free pore space and inversely with the square of the soil volume times the square of the effective specific surface, that is, $p_0^2/U^2(1 - p)^2$. Kozeny (65), on the other hand, has proposed that the value of k varies according to the ratio, $p^3/(1 - p)^2$. It is important to note that a distinct attempt has been made by these investigators to modify Darcy's law in such a way as to evaluate the significance of the amount and nature of the soil-pore space.

✓ *Soil Porosity and Permeability.* The exact relationship between porosity and soil permeability is as yet in the experimental stage. There is sufficient evidence to show that the size, density of packing and hydration of the particles have great effects upon permeability. Lutz (34) has demonstrated that the permeability of clays increases as hydration of the particles decreases. The relative permeabilities of Davidson, Iredell, Putnam and bentonite colloids were found to be approximately 100, 50, 32 and 2, respectively; the relative degrees of hydration of these systems were 0, 10, 35 and 100, respectively. These differences are due to the nature of the colloid, since the SiO_2 -sesquioxide ratio of these systems increases in the same manner as

the hydration. The curves in Figure 32 indicate that the nature of the adsorbed ion also affects hydration and permeability.

Bodman (4) has shown that textural differences exert little influence on permeability at very high apparent densities (volume weights) for soils finer than fine sandy loams. Percolation was exceedingly slow at apparent densities greater than 1.4 to 1.5. A decrease in total porosity, however, does not always give a true picture of the reason for diminished permeability. The content of large pores, that permit ready percolation of water, is the important factor. In terms of the concept of Schumacher (48), the content of non-capillary pores (probably the same as the tension-free pores according to Zunker) largely determines the percolation rate of a uniform soil column.

Burger (11) investigated the effect of forest cover on the properties of the soil and found that the infiltration rate depended upon the air capacity or non-capillary porosity of the soil. (See Figure 38, page 180.) It is usually observed that soils with apparent low permeabilities always have a small content of non-capillary pores. Soils have been found with a non-capillary porosity as low as 2 per cent of the entire soil volume. Such soils were almost completely impervious to water.

In defining non-capillary porosity in terms of the size of pores that do not hold water tightly by capillary forces, it is obvious that the tension under which a soil is drained will affect the amount of non-capillary pores. Under most conditions of experimentation, the soil is saturated by capillarity and allowed to drain under gravity. Recent experimental observations indicate that the permeability of a soil is related to the shape of the pF -moisture curve (3). There are several characteristics of the curve which affect the percolation rate. First, the amount of water which is withdrawn from zero tension to that of the flex point of the curve appears to be closely associated with the rate of water movement through the column. This amount of water represents the pores that have been drained and filled with air. The larger this amount the greater is the permeability. Second, the tension of the flex point in the moisture- pF curve seems to be closely related to permeability. The lower the pF of the flex point, the greater is the rate of percolation. Third, in some soils, there is a tendency towards greater permeability as the slope of the pF curve from zero tension to that of the flex point decreases.

In considering the significance of the first two factors, one of the more permeable soils, Cecil clay, was compressed so that there was 13 per cent more soil per unit volume. The loose soil had a percolation rate of 127 cc. per 10 min.; the rate of the compressed

soil was 5.8 cc. per 10 min. The pF -moisture curves for this soil in the loose and compressed states are shown in Figure 51. It is seen that compression caused several distinct changes in the shape of the curve. The pF of the flex point was raised from 1.65 to 1.80; the amount of water withdrawn from zero tension to that of the flex point was decreased from 12.0 per cent to 5.5 per cent of the total soil volume; total pore space was decreased from 61.8 to 56.8 per cent; and the slope of the curve from zero tension to that of the flex point

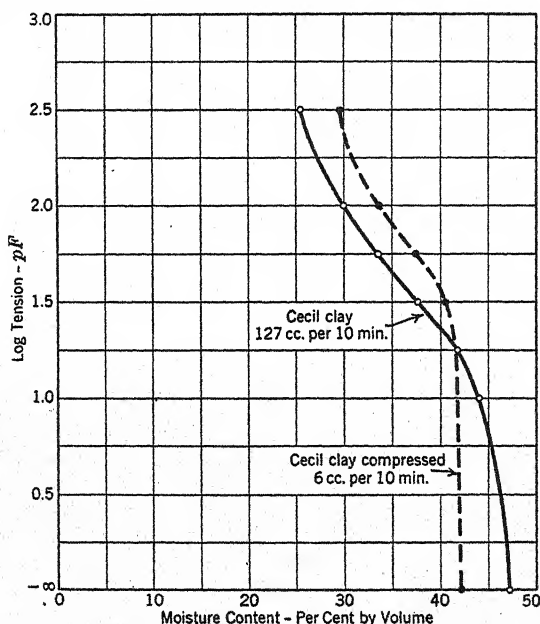


FIG. 51. The effect of compression on the porosity of Cecil clay.

was increased. It should be noted that the slopes of the two curves at tensions above the flex point are practically the same.

Compression effected a change only in the nature of the larger pores. It caused a decrease in the amount and size of the larger pores as evidenced by an increase in the tension of the flex point as well as the slope of the curve below this point. The rather steep slope of the curve from zero tension to that of the flex point indicates a uniformity in size of the pores. In the loose sample, the curve shows a variation in pore sizes up to this point. The increase in the height of the flex point signifies that there has been a decrease in the pore size of that group of pores which are present in the greatest frequency. Compression also has changed the amount of pore space which can

be drained up to the tension of the flex point. In other words, it has decreased the non-capillary porosity. The percentage decrease in non-capillary porosity has been much greater than the change in total porosity. The former was decreased about 54 per cent and the latter

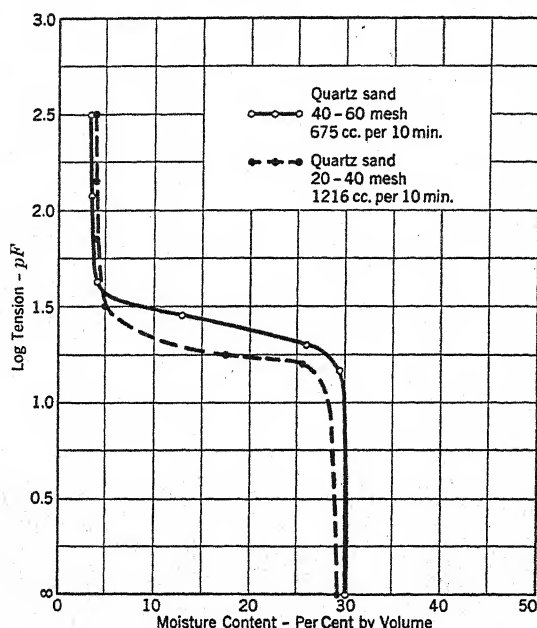


FIG. 52. The pF -moisture curves of two sand separates.

about 8 per cent. Similar effects were obtained with Chenango loam and Davidson clay (Table 25).

In order to show the effect of changing the pF of the flex point upon the permeability of the column, it is interesting to compare the two curves in Figure 52. These curves represent two quartz-sand fractions in which the particle sizes are somewhat uniform. The non-capillary porosity (zero to flex point) is the same; that is, 22 per cent of the total volume. The pF of the flex point of the coarser sample, 20-40 mesh, is 1.25; that of the finer, 40-60 mesh, is 1.50. The percolation rates of these two samples are 1216 and 675 cc. per 10 min., respectively. Similar effects are noticed if one compares the Davidson A-, Davidson B-compressed and Chenango-compressed (Table 25). The non-capillary porosity for each of these soils is 6 per cent. The pF values of the flex points are 1.65, 2.00 and 2.05, respectively; the percolation rates are 17, 4 and 2 cc. per 10 min., respectively.

TABLE 25

PERMEABILITY AND NON-CAPILLARY POROSITY OF VARIOUS SOILS

Soil type	pF of flex point	Non-capillary porosity at flex point, per cent of soil volume	Permeabil- ity, cc./10 min.	Porosity factor
Genesee silt loam—1.....	1.55	14.7	205	9.5
Genesee silt loam—3.....	1.50	13.0	137	8.7
Cecil clay—1.....	1.65	12.0	127	7.3
Cecil clay—2.....	1.75	13.5	136	7.7
Cecil clay— <i>c</i> *.....	1.80	5.5	6	3.1
Davidson clay—A.....	1.65	6.0	17	3.6
Davidson clay—B.....	1.65	8.0	28	4.8
Davidson clay—B— <i>c</i> *.....	2.00	6.0	4	3.0
Chenango loam.....	2.15	11.0	50	5.1
Chenango loam— <i>c</i> *.....	2.05	6.0	2	2.9
Iredell sandy clay loam—A-1...	1.70	9.0	65	5.3
Iredell sandy clay loam—A-2...	1.75	13.0	131	7.4
Iredell sandy clay loam—B....	1.55	9.2	36	5.9
Paulding clay.....	1.60	11.5	93	7.2
Wooster silt loam.....	2.60	10.0	10	3.8
Zeolite.....	1.50	20.0	465	13.3
Quartz sand (40-100 mesh)....	1.55	25.0	850	16.1
Quartz sand (40-60 mesh)....	1.50	22.0	675	14.7
Quartz sand (20-40 mesh)....	1.25	22.0	1216	17.6

* *c* denotes that sample was compressed.

The effect of changing the non-capillary porosity at the same pF value of the flex point is shown by comparing Genesee silt loam-1 with Iredell-B. The pF of the flex point of both soils is 1.55. The Genesee silt loam in natural structure has a non-capillary porosity of 14.7 per cent; that of the Iredell-B is 9.2 per cent. The percolation rate of the former is 205 cc. per 10 min.; that of the latter is 36 cc.

Since the non-capillary porosity (zero to flex point) and the height of the pF of the flex point seemed to be so closely associated with permeability, it was assumed that the percolation rate should vary directly with the amount of the larger pores and inversely with the force or tension required to drain these pores. The ratio of the non-capillary porosity to the pF of the flex should characterize the porosity factor. If the data in Table 25 are plotted to express the percolation rate as a function of the porosity factor, the curve in Figure 53 is

obtained. It is seen that the percolation rate increases exponentially with the porosity factor. It is highly significant that values for such widely varying materials as sand, zeolite, clays and soils in their natural structure should fit the curve so closely. These data justify the contention that soil permeability is dependent upon the non-capillary porosity, if the tension at which this porosity is determined is chosen correctly. Evaluation of the effective pore space from moisture-tension curves appears to offer considerable promise for char-

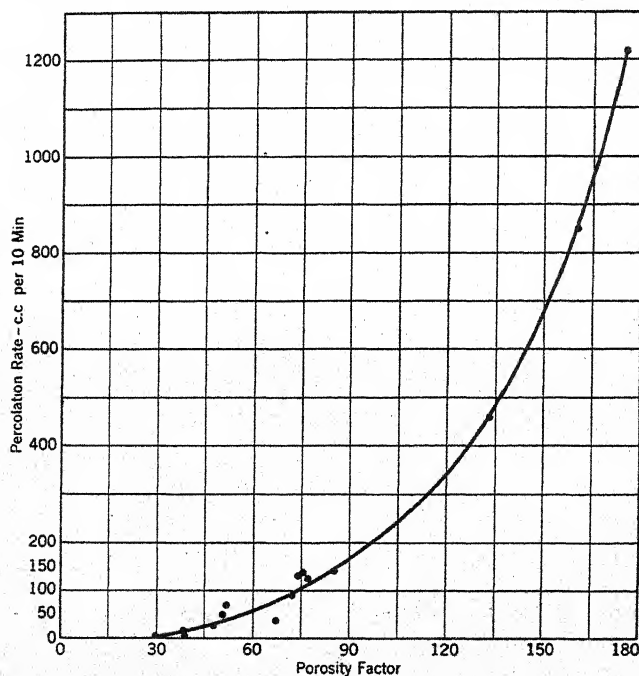


FIG. 53. The relation of permeability to non-capillary porosity.

acterizing the constant for the Darcy equation more accurately in terms of soil properties.

Donat (12) has used the tension-moisture curve to calculate the size distribution of pores in the soil. His method is based upon the well-known relationship of the height of rise of water in capillary tubes. In a given size of capillary, the meniscus will support a certain column of water. If suction is applied to such a tube, it will drain when the tension, in centimeters of water, is equal to the column of water the meniscus will support against gravity. In capillary tubes

the meniscus will maintain a column of water according to the ratio $\frac{\text{circumference of the tube}}{\text{cross-sectional area}}$, or $\frac{U}{F}$. Donat has called attention to the fact that the more the cross section of the pore departs from circular the larger will be the supporting force of the meniscus for a given cross-sectional area. For spherical particles, the supporting force of the meniscus in a loosely packed arrangement is 1.92 times that of a circular cross section; in a closely packed structure it is 2.21 times as large. He has suggested, therefore, that soil pores of a given cross-sectional area will support about twice the tension as equivalent circular pores. The size distribution of pores in several soils is given

TABLE 26

A DISTRIBUTION OF SOIL PORES ACCORDING TO DONAT (12)

(Percentages are in terms of the completely saturated pore space)

Pore size group.....	I	II	III	IV	V
Equivalent diameter of the pores, mm.....	0.02	0.02-0.05	0.05-0.10	0.10-0.20	0.20
Suction tension, cm.....	150	150-60	60-30	30-15	15
pF	2.18	2.18-1.78	1.78-1.48	1.48-1.18	1.18
	per cent	per cent	per cent	per cent	per cent
Quartz sand, 0.5-1.0 mm.. (1)	4.0	4.2	8.8	80.5	2.5
Clayey loam..... (2)	96.0	3.0	0.8	0.2
"Schluff" sand..... (3)	20.0	40.0	35.0	5.0
Field soil (Ackerkrüme)... (4)	63.5	6.0	4.7	5.8	20.0

in Table 26. The tension-moisture curves of these soils are shown in Figure 54.

These results point out that the sand has a high content of large pores as evidenced by the fact that 90 per cent of the water was withdrawn at a tension of 50 cm. of water; 83 per cent of the pores have an equivalent diameter larger than 0.10 mm. The loamy clay, on the other hand, lost only 4 per cent of water at a tension of 150 cm. This soil has but 1.0 per cent of pores larger than 0.05 mm. The "schluff" sand is characterized by a tension-moisture curve that has a gradual slope up to a tension of about 100 cm. of water; there is a rather uniform distribution of pores in this soil. The field soil is granular and a large amount of water was removed at a tension of less than 10 cm. of water. The curve then rises steeply as water is withdrawn from the finer capillaries.

These curves of Donat are similar to many that the author has obtained with approximately the same technique. Permeability of the soil for water is unquestionably a function of the amount of the larger pores. The exact significance of the relative distribution of the pores below a given tension remains to be determined in future investigations.

Profile Characteristics Affecting Permeability. Downward movement of water in soils must take place through different horizons. The porosity and permeability of the various layers may be greatly dif-

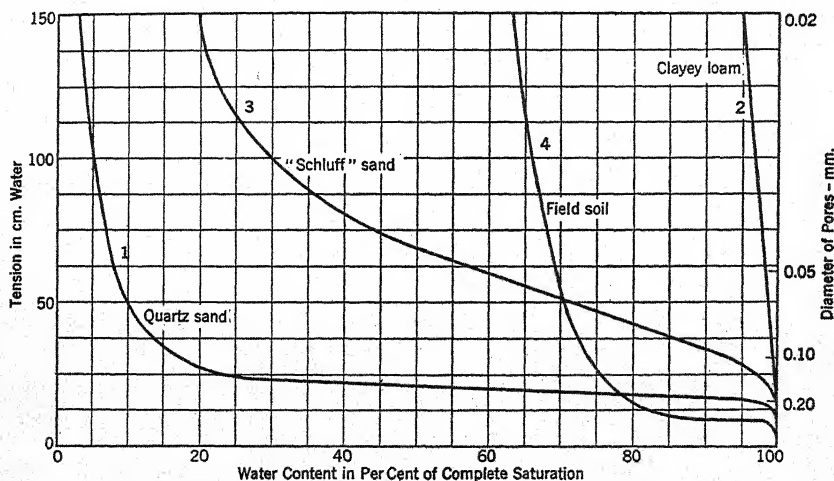


FIG. 54. Donat's (12) tension-moisture curves, showing the relation of water content to the size of pores.

ferent. In a saturated soil, the percolation rate is determined by the permeability of the least pervious horizon. This fact was demonstrated by Wollny (63) who showed that 1 cm. of loam, placed in the middle of a 50-cm. column of sand, reduced the permeability of the column for water about 50 times, depending upon the size of the sand fractions. If one visualizes how a constriction in a funnel changes the rate of flow through it, a rather simple picture of the effect of impervious layers on soil permeability is obtained. The comparison of the non-capillary pore spaces in the Shelby loam and Marshall silt loam in Figure 36, page 163, indicates the importance of a uniform porosity throughout the soil profile on permeability, since the Marshall is about 7 times more permeable than the Shelby.

If an impermeable layer exists in the subsoil, water movement of any consequence through such an horizon can take place only through

cracks and fissures or old root channels and worm holes. It is difficult, if not impossible, to granulate adequately these deeper layers. Deep-rooted plants, such as alfalfa and sweet clover, are effective in producing some channels for water movement if the claypan horizon is not too thick or too compact.

The movement of water in the larger pores may be influenced considerably by the resistance of entrapped soil air. If water that enters the soil leaves some pores in contact with the atmosphere, the soil-air pressure remains unchanged. However, if a rapid entrance of water into the soil entraps appreciable quantities of air, the soil-air pressure will increase and percolation will decrease. This is especially true if a compact claypan below the surface prevents the downward movement of air as well as water.

By way of summary, it can be said that various experimental investigations indicate that the downward movement of water by gravitational forces in natural soils is related (1) to the amount and continuity of the non-capillary pores as determined by soil structure, texture, volume changes and biological channels, (2) to the hydration of the pores and (3) to the resistance of entrapped air.

Movement of Water in Vapor Phase. We have seen that movement of water in the liquid phase ceases below the moisture content at which the capillary permeability is zero. Any movement of water within the soil at these lower moisture contents, therefore, must take place through the vapor phase. Movement in the vapor phase occurs as a result of vapor-pressure differences within the soil. Water always moves from a higher to a lower vapor-pressure region. Differences in vapor pressure may arise from variations in soil moisture or soil temperature. For example, it was shown in Chapter III that the vapor pressure of the soil increased with moisture content as long as the relative humidity of the atmosphere surrounding the soil was less than 100 per cent. Lebedeff (31) has presented data to indicate that the relative humidity of the soil air is always 100 per cent if the moisture content of the soil exceeds the hygroscopic coefficient. This means that movement of water in the vapor phase at moisture contents above the hygroscopic coefficient would depend primarily upon temperature differences between various soil layers or between atmosphere and soil. Undoubtedly, vapor-pressure changes resulting from temperature differences are more important than those arising from moisture variations below the value of the hygroscopic coefficient.

It is a well-known principle that the pressure of a gas increases directly with the temperature if the volume remains the same. Let us apply this principle to the movement of water vapor in soils. Suppose

that we take a 12-inch column of soil with a uniform moisture content somewhere between the field capacity and the hygroscopic coefficient. If evaporation is prevented, there will be no movement of soil moisture. However, let us surround the upper 6 inches of the column with a water jacket and lower the temperature. Such a cooling will lower the vapor pressure of the water molecules in the soil air in the upper half below that in the bottom half of the column. Consequently, water will distill as a vapor upward through the column, causing an increase in the moisture content of the upper 6 inches.

Lebedeff (31) has attached considerable significance to the movement of water in the vapor phase. He has reported that about 72 mm. of water per year condenses from the air in a chernozem soil (at Odessa, U.S.S.R.). The total rainfall averages 400 to 450 mm. For example, he found that the moisture content of the upper layer of soil at a depth of one centimeter increased from 1.99 per cent at 4 P.M. on July 27, of a dry year, to 5.26 per cent at 4 A.M. on July 28. The relative humidity of the atmosphere varied between 50 and 92 per cent. Part of this moisture was obtained by condensation from the atmosphere and part from the movement of water vapor upward from lower horizons. The intensity of condensation from the atmosphere increases the greater the difference between the absolute humidity of the air and the vapor pressure of the soil. The amount of movement upward increases with the difference in temperature between the upper and lower layers.

Lebedeff has called attention to the fact that a cool spell during the summer not accompanied by rain, which is followed by a hot period, may result in considerable losses of water from the deeper layers. As the surface is cooled, the vapor pressure is lowered and upward water-vapor movement occurs. Meteorological observations on moisture changes at the University of Odessa during the winter of 1914-1915 have revealed the very interesting data in Table 27. It is noted that 66.2 mm. of water in the form of vapor moved from the deeper horizons to the surface during the winter. During the winter of 1922-1923, at Rostov, 66 mm. were found to have moved upward from October 7 to March 6. These moisture movements have been explained on the basis of differences in vapor pressure resulting from warm subsurface and cool surface layers. In the summer time, a warm surface and cooler subsurface horizons cause a downward movement of moisture. Although these experiments of Lebedeff have not been duplicated by other investigators, they emphasize the possibility of considerable movement of soil water in the vapor phase under certain conditions.

TABLE 27

WATER RELATIONSHIPS IN SOIL AT ODESSA, U.S.S.R., 1914-1915 (LEBEDEFF, 31)

Observation	From October 26 to December 16	From December 16 to March 1	From October 26 to March 1
	mm.	mm.	mm.
1. Observed increase of water in soil.	70.9	109.7	180.6
2. Amount of water evaporated from the soil.....	23.6	31.5	55.1
3. Actual increase of water in the soil (the sum of 1 and 2).....	94.5	141.2	235.7
Amount of precipitation.....	49.8	119.7	169.5
Water moved from the deeper horizons in the form of vapor...	44.7	21.5	66.2

Evaporation of Water from Soil. Evaporation of water from soils has been the subject of much controversy for a long time. This is especially true from the point of view of the practical significance of water losses by evaporation and methods for controlling or minimizing such losses. Almost everyone is familiar with the fact that the soil dries out after a rain, at least in the immediate surface. Moreover, it is recognized that this drying-out process depends upon weather conditions and the nature of the soil. A long-debated question has been concerned with the amount of water that evaporates from soils after the surface has become relatively dry. In this discussion, we shall call attention to the various factors that affect the evaporation of water from soils and then analyze the existing information concerning the rate of evaporation of soil moisture from depths below the immediate surface.

METEOROLOGICAL FACTORS AFFECTING EVAPORATION. Evaporation of water from any source can occur only when the atmosphere in contact with the water is not saturated with water vapor, if both air and water have the same temperature. If a pan of water is placed in a closed container, evaporation will proceed until the air is saturated with water at that particular temperature. If the temperature is then lowered, part of the water-vapor molecules will condense to form liquid water. Consequently, it is obvious that any meteorological effect that tends to increase the vapor-pressure gradient away from the soil will increase evaporation.

The effect of temperature on evaporation is rather obvious. A rise in temperature increases the vapor pressure of water and the rate

of evaporation is enlarged. Evaporation is greater, therefore, in the summer months than during any other season. An analysis of the evaporation data from 243 monthly records from 29 meteorological stations throughout the United States has indicated that evaporation from a free-water surface varies approximately with the square of the mean monthly temperature in Fahrenheit degrees (2). Although this value does not take the effect of wind velocities into consideration, it illustrates the importance of air temperatures in the evaporation process. The direct rays of the sun also speed up evaporation by increasing the temperature of the soil surface.

The relative humidity of the air is an extremely important factor in evaporation. It is recognized that soils do not dry out very rapidly during periods in which the air is nearly saturated with water vapor. On the other hand, evaporation of water in the arid regions is a rapid process. The effect of the humidity of the atmosphere on evaporation and the effectiveness of rainfall have been used by Meyer (23) as a basis for classifying soil-climate relationships. He formulated the concept that the effective soil moisture was proportional to the precipitation divided by the saturation deficit of the air. This quotient, called N. S. Q. (Niederschlag-Sättigungsdeficit Quotient), has been employed by Jenny (23) to classify the climatic soil types of the United States. The influence of the humidity of the air is well illustrated by the data of Masure (37), which show that evaporation at 17.0 to 17.5° C. increased from 0.25 to 0.93 mm. as the humidity decreased from 91 to 75 per cent.

The wind velocity also has a great effect upon evaporation. Wind movement exerts a continual disturbance of the air immediately above the soil. Air that may have become fairly well saturated with water vapor is displaced by drier currents. This speeds up the evaporation process. The greater the velocity of the wind, the more the evaporation. For example, Wollny cited data by Hensele which showed that a 12-mile-per-hour wind caused 7.8 gm. of water to evaporate from 100 sq. cm. of a granular loam as compared with 0.3 gm. in still air. Hensele also pointed out that a warm, dry wind brought about the greatest evaporation at a given velocity. Common experience has demonstrated the serious drying effects of hot winds, especially in the semi-arid regions.

SOIL FACTORS AFFECTING EVAPORATION. The degree of saturation with moisture is the most important soil factor affecting the amount of evaporation. The results of Fisher (15) and Keen (27) illustrate this point rather clearly. Both investigators have shown that the rate of evaporation is practically constant at high moisture contents.

Fisher has attributed this constant rate to the fact that water essentially is evaporating from a free-water surface. He has compared the evaporation from soil with that from wool fibers and has suggested four distinct portions to the rate of evaporation-moisture curve, starting with a wet soil:

1. *Evaporation rate constant:* Evaporation proceeds at a constant rate as the moisture content decreases, owing to the similarity to a free-water surface.

2. *Evaporation rate proportional to the water content:* This proportionality is due to the ready movement of water from the larger pore spaces.

3. *Evaporation rate proportional to the rate at which water comes to the surface from the inner capillaries.*

4. *Evaporation rate due to hygroscopic moisture:* This rate is very slow since water must travel in the vapor phase from within the soil to the surface.

Fisher was of the opinion that the moisture contents at which these rates changed were characteristic for each soil and were some function of surface.

Keen, however, demonstrated that the technique of measuring evaporation affected the nature of the curves considerably. He found that the rate of evaporation curves was quite similar to that of vapor-pressure curves and came to the conclusion that evaporation is dependent upon the available surface and the vapor pressure in the moist soil. Both these values decrease with diminishing moisture content.

The factor of the depth to the water table is closely related to the degree of saturation of the soil with water. That is, evaporation of water from a drained soil, that does not have a water table relatively close to the surface, is entirely different from that from a soil in contact with a free-water table. Eser (13) demonstrated, as early as 1884, that evaporation from soils in contact with ground water was 2 to 4 times greater than from drained soils. (See Table 28.) King (30) measured the evaporation of water from various soils in relation to the depth of tillage. Soil columns were placed in contact with a free-water surface and evaporation allowed to proceed for 100 days. These classical data are shown in Table 29. Most water was lost from the clay soil. Under these conditions, mulching reduced evaporation losses as much as 63 per cent.

The evaporation experiments of King were interpreted as emphasizing the importance of a dry, loose layer of soil on the surface in order to break capillary connections with the immediate surface.

TABLE 28

EVAPORATION FROM AN EXPERIMENT FIELD SOIL WITH DIFFERENT COVERINGS
(ESER, 13)

Type of cover	Water lost per 1000 sq. cm. from July 12 to August 12, 1883					
	Bare	0.5 cm. Chopped straw	5 cm. Chopped straw	5 cm. Beech leaves	5 cm. Pine needles	5 cm. Fir needles
Water lost—gm...	5739	2392	571	630	878	621

The so-called dust mulch was advocated as an essential practice in conserving water by reducing evaporation. Recent data by Veihmeyer (55) and others, however, have shown that evaporation losses are confined to relatively shallow depths and that most of the water that is lost through evaporation occurs before the moisture content is lowered sufficiently to create a "dust mulch" on the surface. These concepts are not in contradiction to King's findings since they refer to soil conditions where capillary rise from a near-by water table does not exist.

TABLE 29

EFFECTIVENESS OF SOIL MULCHES FOR CONTROLLING EVAPORATION (KING, 30)

Type of mulch	Inches of water lost in 100 days		
	Black marsh soil	Sandy loam	Virgin clay loam
None.....	5.193	6.548	21.31
1 in. deep.....	3.12	3.300	11.13
2 in. deep.....	2.384	2.996	8.652
3 in. deep.....	2.265	2.539	7.852
4 in. deep.....	2.230	2.785	7.805

If evaporation is analyzed from the standpoint of capillary movement in relation to the tension forces in the soil, it is seen, as discussed on page 216, that capillary conductivity approaches zero in the vicinity of the field capacity. When one considers that most soils become plastic at moisture contents near the moisture equivalent, one sees readily that stirring the soil to form a "dust mulch" can be done only

after most of the water that will evaporate without stirring has been lost.

Particle size and the state of aggregation exert a significant influence upon evaporation, especially when the soil is wet. Eser (13) studied the amount of evaporation of water from sieved soils that took place per 1000 sq. cm. of surface from June 2 to July 3, 1883. If the amount evaporating from the fractions smaller than 0.071 mm. in diameter is taken as 100 per cent, the loss from the 0.25 to 0.5 mm. fraction is 81.1 per cent and that from the 1.0 to 2.0 mm. separates is 22.2 per cent. The greater evaporation from the finer fractions is undoubtedly due to their greater capillary capacities.

The two factors of soil color and direction of exposure affect evaporation primarily as they influence soil temperature. Black soils absorb more heat from the sun than light-colored soils; they also lose more water through evaporation; they gain more water by condensation at night. Evaporation is greater on south than on north slopes; on a south exposure it increases with the degree of slope.

The uses of artificial mulches greatly retards evaporation. The data in Table 28 point out the beneficial effects of organic mulches. Such mulches protect the soil from the direct rays of the sun and from wind currents. Consequently, the soil is kept cool and the vapor pressure of the air within the mulch is more nearly the same as that within the soil air. Artificial mulches are used rather frequently in orcharding and gardening.

SOIL-MOISTURE EQUILIBRIUM POINTS

Since there are several equilibrium points that are used rather extensively to express soil-water relationships, it is essential that some of the salient features of these values be mentioned briefly.

Hygroscopic Coefficient. The hygroscopic coefficient has been discussed in detail in Chapter III. It represents the amount of water that is adsorbed on the surface of the particles from an atmosphere of water vapor of known relative humidity. It is an index of the surface activity of the soil, if it is employed carefully. The Mitscherlich value for hygroscopicity, which is used extensively in Europe, is determined over 10 per cent H_2SO_4 , which gives a relative humidity of 94.3 per cent at 25° C.; this is equivalent to a pF of 5. The investigators at Rothamsted employ 47 per cent H_2SO_4 , which produces a relative humidity of 50 per cent; the pF is about 6. In the soils investigations of the United States Department of Agriculture, 3.3 per cent H_2SO_4 , giving a relative humidity of about 98.2 per

cent, and 30 per cent H_2SO_4 , which produces a relative humidity of 74.9 per cent, are used. The former corresponds to a pF of 4.5 and the latter to a pF of about 5.6. Thirty per cent H_2SO_4 is employed to provide information on the energy of adsorption of water by a given colloid; the amount of adsorption over 3.3 per cent H_2SO_4 is supposed to be an index of the total amount of surface.

It should be remembered that these so-called equilibrium values represent only certain points on the vapor pressure-moisture curve. The attractive forces in the surface of the soil colloidal material are primarily responsible for the adsorption of water.

Wilting Point or Wilting Percentage. The wilting point refers to that soil-moisture content at which soil cannot supply water at a sufficient rate to maintain turgor, and the plant permanently wilts. Briggs and Shantz introduced the concept of the wilting point as an important soil-moisture constant. Plants were grown in pots, in which evaporation was prevented, and the moisture content of the soil determined when the plant permanently wilted. They contended that all plants reduce the moisture content of soils to about the same extent when wilting occurs. On this basis, they proposed that the wilting coefficient could be calculated from the moisture equivalent by dividing the latter by the factor, 1.84.

Numerous investigations since the original work of Briggs and Shantz have shown that this factor cannot be used for all soils. Veihmeyer and Hendrickson (57), however, have confirmed the original contention of Briggs and Shantz that all plants reduce the moisture content of a given soil to about the same value before wilting occurs. Sunflower plants are now used most frequently as the test plant for measuring the wilting point of a soil. All indirect methods are unreliable.

The tension of the soil water when permanent wilting occurs is at a pF of about 4.2. Water is probably held as a thin film around the particles at this tension. At least, any water wedges at the point of contact of the particles must be very small. Movement of water within the soil takes place in the vapor phase, since the capillary conductivity is zero. It is highly probable that the moisture content of the soil adjacent to the roots is much lower than the average water content of the pot because of the immobility of the soil water at this point. Plants may extract water to values below the wilting point insofar as their root system comes into direct contact with moisture films.

Field Capacity. Veihmeyer and Hendrickson (58) have defined the field capacity as "the amount of water held in the soil after the excess gravitational water has drained away and after the rate of

downward movement of water has materially decreased." This value has been referred to as the field-carrying capacity, normal field capacity, normal moisture capacity and capillary capacity. Shaw (49) showed that the addition of 6 in. of water to the surface of a column of dry soil 36 in. long resulted in a uniform moisture content to a depth of 24 inches. He called the constant moisture content throughout this depth the normal moisture capacity. Similar results were reported by Alway and McDole (1), who pointed out that soils drain to a moisture content about 2.5 times that of the hygroscopic coefficient.

Alway and McDole found that the field capacity of sands was higher than the moisture equivalent but that the value for loams was about the same as the moisture equivalent. Shaw came to nearly the same conclusion. Veihmeyer and Hendrickson have also substantiated the fact that the moisture equivalent of sands is below the field capacity. Water is held at a tension equivalent to a pF of about 3.0 to 3.2 at field capacity. Moore (38) has shown that capillary conductivity approaches zero at this tension.

✓ Since soils normally drain to a moisture content of the field capacity and since plants permanently wilt at the wilting point, it is seen that the amount of water available to plants is represented by the difference between these two equilibrium values. Consequently, the nature of the tension-moisture curve of a given soil will give a fairly good measure of the availability of soil water. If the curve is steep between these two points, there will be little available water and it will be difficult to grow plants satisfactorily. Water will be held too tenaciously to be removed by the plant roots at a rate rapid enough for growth. If the curve is gently sloping between the field capacity and the wilting point, water will be available.

The question has often arisen as to the effects of organic matter on water absorption and availability. Many people have the idea that since organic matter has such a high water-holding capacity its addition to the soil should increase the amount of water available to plants. A fairly good picture of these effects is found in the data of Feustel and Byers (14) in Table 30. These results show that organic matter possesses a high water-holding capacity. Likewise, it has a high wilting percentage. It is significant to note that a 50-50 mixture of clay and peat did not appreciably increase the amount of available water, although the water-holding capacity was increased. Quartz sand, however, was benefited by additions of peat. Feustel and Byers have stated that "such mixtures with soil were capable of absorbing from 40 to 50 per cent more moisture than the soil alone, but the increased evaporation rate and the greater content of moisture at the wilting

point largely counteracted the initially higher moisture-holding capacity." On the basis of their results, they have recommended that peat should not be used for the sole purpose of conserving a supply of available soil moisture except in the case of highly decomposed types, on sandy soils. It should be mentioned, however, that surface additions of organic matter may exert beneficial effects on the available moisture in soils, apart from its moisture-holding properties, by causing a greater infiltration of water.

TABLE 30

THE RELATION OF ORGANIC MATTER TO THE WATER-HOLDING CAPACITY AND AVAILABILITY OF WATER (FEUSTEL AND BYERS, 14)

	Maximum moisture- holding capacity, per cent	Moisture equiv- alent, per cent	Wilting per- centage, per cent	Avail- able water, per cent
Clay loam soil.....	44.3	20.2	7.1	13.1
Quartz sand.....	28.3	1.4	0.57	0.83
Moss peat.....	1,057	166	82.3	83.7
Sedge peat.....	374	112	60.8	51.2
Reed peat.....	289	110	70.7	39.3
$\frac{1}{2}$ clay soil + $\frac{1}{2}$ moss peat.....	114	31	14.5	16.5
$\frac{4}{5}$ clay soil + $\frac{1}{5}$ moss peat.....	57.3	21.6	8.5	13.1
$\frac{1}{2}$ quartz sand + $\frac{1}{2}$ moss peat....	89.1	12.7	5.2	7.5
$\frac{4}{5}$ quartz sand + $\frac{1}{5}$ moss peat....	47.8	5.6	1.8	3.8

The Moisture Equivalent. The moisture equivalent is one of the most frequently used determinations for characterizing the moisture relations of soils. It was introduced in 1907 by Briggs and McLane (8), who centrifuged saturated soils in perforated cups at a speed equivalent to a force of 1000 times that of gravity. They considered this force to remove most of the water that is held in the larger pores. Later, Briggs and Shantz (9) used the moisture equivalent as an indirect measure of the wilting point. Veihmeyer and Hendrickson (58) have shown that the moisture equivalent gives a fairly reliable measure of the field capacity of fine-textured soils. It also has been used by several investigators as a means of expressing soil texture.

Fisher (16) introduced the idea that the moisture equivalent measured the water held in the small pores as well as that imbibed by the soil colloids. He suggested that soils be saturated with xylene and

centrifuged to give the amount of pore-space water. It was assumed that the colloids would not imbibe xylene. Therefore, the difference between the moisture and xylene equivalents would be an index of the imbibitional water. This concept offers many possibilities for readily determining the relative hydration of various soils.

The reader is referred to the papers of Veihmeyer, Oskerkowsky and Tester (59) and of Joseph and Martin (24) for a comprehensive discussion of the various experimental factors affecting the value of the moisture equivalent. The former have recommended that air-dried samples, that have been crushed to pass a 2-mm. sieve and that have soaked 24 hours, be used for making the determinations. The latter have called attention to the fact that highly dispersed and impermeable soils may become water logged and give high results.

The force with which water is held at the moisture equivalent is equal to a pF of about 2.70. According to Moore (38), capillary conductivity is approaching zero near this moisture content. Water movement is very slow at moisture contents below the moisture equivalent.

✓ DETERMINING SOIL MOISTURE ✓

Soil moisture is usually determined by taking a sample of soil from the field and driving off the water at 105 or 110° C. The percentage of water held by the soil on a dry basis is the moisture content. Thus, it is expressed on a percentage by weight basis. Although the calculation of the moisture percentage on the weight basis is very simple, it is doubtful if this method of expressing the amount of water in the soil is the most advantageous. From the standpoint of degree of saturation of pore space with moisture, the water capacity is more readily visualized if expressed on the volume basis. Most hydrologists like to speak in terms of inches of water per given depth of soil. If soil moisture were calculated on a volume basis, it would be easy to convert to inches per acre or inches per foot.

There are several advantages in using volume percentages for denoting soil moisture. In the first place, it helps to picture the relationship of soil to air and water. A given soil may be visualized as possessing a certain total porosity which is partly filled with water and partly with air. Moreover, it is possible for two soils to have the same moisture content on a percentage by weight basis but not on the volume basis. Second, water retention in the soil mass is within a given volume, not a given weight. The roots of vegetation obtain water from the volume of soil through which they penetrate.

It seems, therefore, that more emphasis should be placed on at-

tempting to express soil moisture on the volume basis. This will necessitate obtaining definite volumes of soil during the sampling process, which will introduce sources of error not present in the usual techniques. Nevertheless, the information that is gathered will be of greater value.

✓ Various attempts have been made to determine soil moisture without resorting to the laborious sampling, weighing and drying techniques. ✓ Elimination of the drying process has been attempted (1) by measuring the change in concentration of alcohol when placed in contact with the moist soil, (2) by determining the amount of gas evolved when calcium carbide is mixed with the moist soil and (3) by measuring the heat evolved when the moist soil is placed in a concentrated sulfuric acid solution. These methods are not sufficiently reliable to warrant their use for accurate soil-moisture measurements. In a majority of the cases, they will give approximate values.

Continued attempts to measure moisture changes in situ have been made with varying success, although considerable progress has been achieved in recent years. These attempts may be classified into two distinct types of procedure: (1) electrical methods and (2) techniques based upon the so-called suction force of soils or capillary tension.

Electrical Methods. In 1897, Whitney, Gardner and Briggs (61) proposed an electrical conductivity method for measuring soil moisture in the field. Electrodes were placed in the soil and the conductivity was measured and interpreted on the basis that any changes in electrical conductivity were brought about by varying amounts of water between the electrodes. It was soon observed, however, that small changes in the salt content of the soil solution affected the conductivity more than the amount of water that was present. In light of this fact, measuring soil moisture by means of electrical conductivity has never proved successful. Multiple electrodes have been used but have not proved entirely satisfactory (35).

It has recently been shown that heat conductivity in soils can be used as an index of soil moisture (50). Use is made of the principle of the increase in resistance of a wire conductor with increase in temperature to measure the changes in heat conductivity of the soil-water system. An element consisting of enameled-copper wire, wound onto a piece of glass tubing and having a resistance of 7 ohms at 0° C., is placed into the soil. It is connected through a Wheatstone bridge arrangement with a microammeter. Starting with a balanced bridge, a current of 0.4 ampere is passed through the bridge for one minute, at which time the current going through the microammeter is read. The magnitude of the current going through the microammeter reflects the

rise in temperature and increase in resistance of the element in the soil. The ability of the soil to conduct heat away from the element determines the temperature rise. Since the heat conductivity of the soil varies with the moisture content, the reading on the microammeter is a measure of soil moisture.

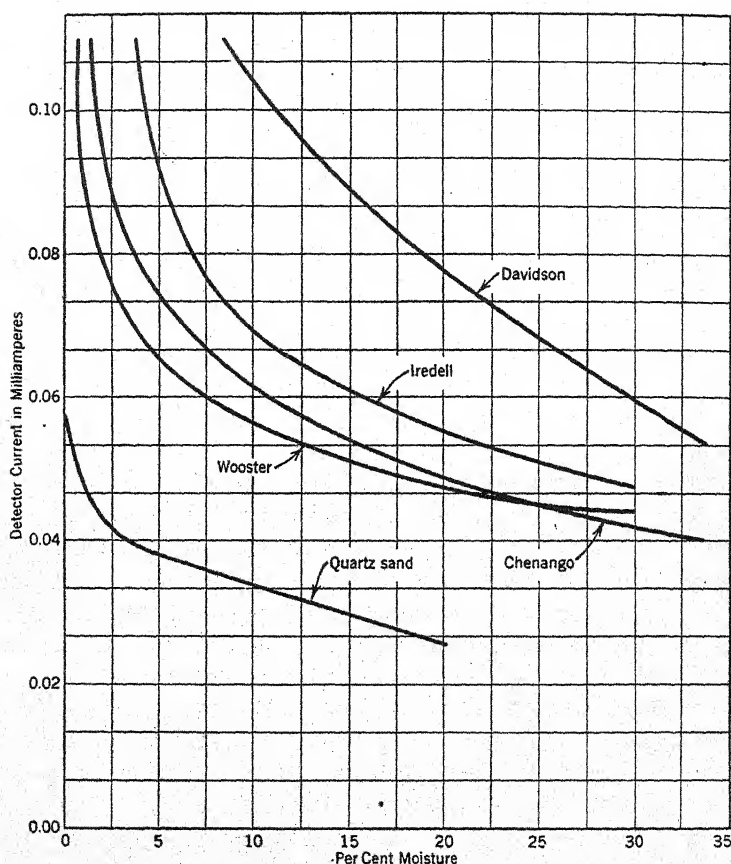


FIG. 55. The relation of heat conductivity to the moisture content of soils. (A high detector current signifies a low heat conductivity.)

Readings with this apparatus are not affected by the presence of electrolytes or changes in soil temperature. Readings can be made over the entire moisture range from saturation to the air-dried state. Results obtained by this technique for several widely different soils are shown in Figure 55.

✓ The relationship between heat conductivity and moisture content

is distinct for each soil. The decrease in heat conductivity (an increase in the detector current) with moisture is rather gradual in the high moisture range. The sensitivity of this method is greatest at moisture contents below the moisture equivalent. Duplication of readings at any moisture content is within about one per cent variation. Soils in the natural and artificially packed state have been used to ascertain the values of this technique for measuring soil moisture. Better results have been obtained with the soils in their natural structure. Although this electrothermal method is as yet in its experimental stages, it offers many possibilities for solving many of the perplexing soil-moisture problems.

Fletcher (17) has recently adapted the dielectric method for measuring moisture to soils. A special condenser is placed in the soil and allowed to attain equilibrium with the soil water. Readings are then taken by means of a suitable capacitor. Readings are affected by the amount of moisture, the colloid content and the salt concentration. Inasmuch as salt concentrations in excess of 0.1 gm. per 100 cc. give constant readings at the same moisture content, the salt error is usually not significant, especially in arid soils. It is possible to bury the condenser in a given soil and follow the moisture changes, once the relationship between moisture content and the dielectric constant for this soil has been established.

Use of Tensiometers. Since 1935, the principle of capillary tension has been used by many investigators to measure moisture changes in the soil. Porous-clay cups or tensiometers, filled with water and attached to a mercury manometer or recording device, are placed in the soil. Water moves into or out of the cup as the soil wets or dries. As discussed on page 206, the use of tensiometers is limited to moisture contents above the moisture equivalent. They cannot be employed at tensions greater than one atmosphere.

Richards and his associates (43, 44) have employed tensiometers to follow moisture changes throughout the soil column with rather good success. That is, the tensiometer indicated whether moisture moved in any particular direction even though the amounts of water involved in such movement were not ascertained. Russell and Richards * installed tensiometers in field soils and determined the moisture content of the soil adjacent to the porous cups at different capillary tensions. Typical results obtained with this technique are shown in Figure 56.

Although considerable variation exists between the moisture con-

* Data as yet unpublished.

tents at a given pF , it is seen that each soil has a characteristic curve. Hysteresis effects and imperfections in the sampling method probably account for the major part of these variations.

Tensiometers undoubtedly will prove to be valuable tools in studying relative changes in soil moisture at contents higher than field capacity. The fact that they cannot be used within the moisture range from the moisture equivalent to the wilting percentage, as well as the

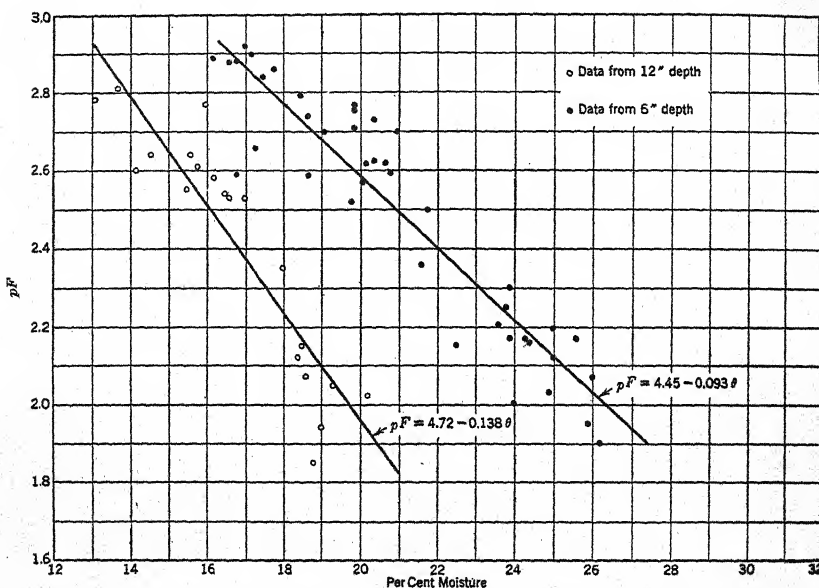


FIG. 56. The use of tensiometers for measuring the moisture content of soils. (Courtesy of Russell and Richards. Data as yet unpublished.)

problem of overcoming hysteresis effects, are definite limitations in their usefulness in the field.

Water-supplying Power of the Soil. In 1920, Livingston and Koketsu † devised the "soil-point method" for measuring the ability of soils to supply water to plants. This method makes use of small porous, porcelain cones that have a known small area of unglazed surface that contacts the soil. The cones are dried, weighed and then inserted into the soil for a given period. Two-hour exposures were used in the original work; in later investigations, a one-hour exposure was employed. They found that the soil points removed about 85 mg. of water during 2 hours when the moisture content of the soil

† *Soil Science*, 9 : 469-485, 1920.

was near the wilting coefficient. (This fact suggests capillary movement although it is possible that the "soil point" takes up water in immediate contact with the soil without capillary movement to it.)

Wilson and Livingston* later introduced the term "soil-moisture index" to denote the water-supplying ability of the soil under different plants. A value of 100 was taken to express a water-supplying power of 8 mg. per sq. cm. of soil during one hour. This value was used because it represented the minimum rate that the soil could supply water to the plant before drought injury occurred. The soil-moisture index varied from 52 for tall oat grass to 29 for red fescue. These results show that red fescue maintains growth on much drier soils than tall oats grass.

The use of soil points requires that the absorbing surface of unglazed porcelain be in good contact with the soil and that the cone absorb water as fast as the soil can release it. From a theoretical point of view, the porcelain cone possesses a certain "capillary pull" as a result of the tension of the water menisci in the pores. The lower the tension of the soil water, the greater will be the water intake by the cone. The method offers considerable possibilities for obtaining a close approximation of the water-supplying power of the soil. The need for irrigation may be predicted with them.

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* *Plant Physiol.*, 7 : 1-35, 1932.

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CHAPTER VII

SOIL AIR

In the discussion of soil structure, attention was called to the fact that a given volume of soil is composed of solid soil particles and pore space and that the pore space is partly filled with water and partly with air. Air relationships, therefore, are closely dependent upon the amount of water in the soil-pore space. When one thinks of the effect of soil air on the properties of soil, especially in relation to plant growth, the amount of air, its composition and method of renewal appear to be most important.

AIR CAPACITY OF SOILS

Kopecky (7) has defined air capacity "as the quantity of air in the soil which remains there, after the soil has been saturated with water to the point of absolute water capacity" (water retained in the soil 24 hours after it has been saturated and allowed to drain). In other words, the air capacity is usually the same as the volume of the non-capillary pores.

Relation of Air Capacity to Texture and Structure. Factors affecting the non-capillary porosity of soils have been discussed in Chapter V. The data in Table 19, page 161, clearly show the relation of air capacity to texture. Sands and mucks contain many large pores; they have high non-capillary porosities. It is an accepted fact that most sandy soils have too much aeration at the expense of an adequate water-holding capacity. Clays, on the other hand, have rather low air capacities. Sufficient aeration is one of the major problems for good plant growth on clay soils. If the individual clay particles are closely packed together, there are no large pores for aeration. The effect of compaction on the decrease in non-capillary porosity is shown in Table 25. Compression of the Cecil clay decreased the air capacity (non-capillary porosity) from 13.5 to 5.5 per cent; the air capacity of the granular Davidson clay was diminished from 8.0 to 6.0 per cent; and the air capacity of the Chenango loam was lowered from 11.0 to 6.0 per cent by compression.

If a clay soil is well granulated, however, there will be a sufficient number of large pores between the granules to give an air capacity that will be satisfactory for plant growth. Doiarenko showed that the size of the granule is an important factor in regulating the non-capillary porosity of soils. For example, in Table 20, it is seen that granules smaller than 0.5 mm. in diameter give an air capacity of only 2.7 per cent. The air capacity increases to 29.6 per cent with granules the size of fine gravel (1.0–2.0 mm.). There is a limit, however, to which the air capacity can be raised as the size of the granules or clods is increased. Yoder (18), in preparing various types of seedbeds for cotton by regulating the size of clods, found that clods smaller than $\frac{1}{16}$ in. gave an air capacity of 29.6 per cent by volume. The air capacity increased to a maximum of 42.4 per cent with separates ranging from $\frac{1}{8}$ to $\frac{1}{4}$ in. in diameter and then decreased to 36.1 per cent with clods larger than 4 inches. It is interesting to note that the values of non-capillary porosity for the $\frac{1}{8}$ -in. and the $\frac{1}{8}$ - to $\frac{1}{4}$ -in. fractions are of the same order of magnitude as those reported by Doiarenko for similar fractions of an entirely different soil. Undoubtedly, the state of aggregation of soils is quite important in determining the air capacity.

The air capacity of heavy soils can be increased by the addition of organic matter, cinders or sand. Organic-matter additions in the form of residues, manure or peat appear to be the most practical means of improving the air relations of clay soils. For example, recent studies at the Ohio State University on the pore-space relations of soils receiving various treatments have provided the following interesting data: check soil, 47 per cent total pore space, 4 inches of cinders mixed within the plow layer, 54 per cent; 50 tons of manure spaded under, 70 per cent; and a mixture of manure and cinders, 71.5 per cent total porosity. The air capacities of these soils were 15.5, 27.0, 22.5 and 26.5 per cent, respectively. It is seen that cinders increase the air capacity but decrease the water-holding capacity. Manure, on the other hand, increases both the total porosity and air capacity.

The non-capillary porosity of the surface soil is a very dynamic property. It changes with weather conditions and the frequency of tillage. Tillage operations usually increase the total porosity and air capacity of a compact soil. On the other hand, extreme pulverization and rolling may lead to a decrease in non-capillary porosity. The impact of raindrops on a granular, well-aerated soil tends to disperse the soil particles and cause a compaction of the immediate surface. There is a decrease in porosity, especially the larger or non-capillary pores. Wollny (15) was the first to show how great this decrease in non-capillary porosity can be. His data have been discussed on

page 179. (See Table 22.) By way of repetition, it is important to stress that the most serious effect of the deterioration of the structure of the surface soil is the decrease in the air capacity. This decrease in the content of non-capillary pores is responsible for poorer aeration, reduced bacterial activity and a lower infiltration capacity for precipitated water.

Under special conditions of excessive packing, when the soil is wet, the air capacity may be reduced to almost zero. For example, frequent tramping of greenhouse soils between the rows of growing tomato plants has been found to lower the air capacity from about 16 per cent by volume to only 2 per cent. This decrease occurred over about one-third of the total area of the greenhouse. Similar effects may be produced on farm lands by the tramping of livestock and the wheels of farm machinery.

The air capacity of soils also may be increased by cracking due to shrinkage, by the roots of vegetation and by the burrows of the animal life in the soil. The classical example of such beneficial effects is the work of Wollny (16) on earthworm activities in soils. He showed that earthworm burrows raised the air capacity of a loam soil from 8.9 to 31.2 per cent. The total porosity was increased less than 3 per cent on the basis of the entire soil volume.

Air Capacity Requirements for Plants. Unfortunately, in spite of the fact that much has been said about the need for adequate aeration for plant growth, there is a lack of information on the exact air requirements of various plants. Kopecky (7) has proposed the following air-capacity ranges for the optimum growth of several farm crops: Sudan grass, 6 to 10 per cent; wheat and oats, 10 to 15 per cent; barley and sugar beets, 15 to 20 per cent. It is fairly well established from practical experience that these crops are arranged in the correct order as far as their relative needs for aeration are concerned. Sugar beets and potatoes are two crops that will not grow satisfactorily on poorly aerated soils. Recent experiences with the growth of sugar beets on the heavy clay soils of northwestern Ohio definitely confirm the high requirements of beets for aeration.

Yoder (18) obtained the highest yield of seed cotton on artificially prepared seedbeds when the non-capillary porosity was over 30 per cent. Clod separates ranging from $\frac{1}{8}$ to $\frac{1}{4}$ inch in diameter (approximately 3 to 6 mm.) gave the best results. Kvasnikov's (13) investigations have shown that millet and spring wheat produce best on seedbeds that are composed of clods from 1 to 3 mm. in diameter. (The non-capillary porosity probably varied between 30 and 40 per cent on the basis of Doiarenko's and Yoder's work). Only about

one-fourth as much grain was produced when the soil clods or granules were smaller than 0.5 mm. (non-capillary porosity undoubtedly less than 10 per cent).

These scattered experimental investigations point out the need of further research in this important phase of soil-aeration problems. There seems to be ample justification for assuming that the non-capillary porosity of the seedbed is an important factor in successful crop production, at least within certain limits.

Relations between Air Capacity and Drainage. In discussing soil-water movement in Chapter VI, data were presented to show that the percolation of water through a soil column was definitely a function of the amount and size of the larger pores. The interpretation of this relationship to drainage problems would imply that the ability of soils to drain naturally as well as their need for artificial drainage are correlated with the non-capillary porosity or air capacity. Kopecky (7) has recommended that all soils that do not have an air capacity greater than 10 per cent by volume need artificial drainage. His results on the relation between the air capacity and the need for drainage are given in Table 31. These data suggest that the increase in air capacity

TABLE 31

RELATION BETWEEN AIR AND WATER CAPACITY AND DRAINAGE (KOPECKY, 7)

Type of soil	Particles smaller than 0.01 mm.	Water porosity	Total capacity	Air capacity	Need for drainage
	per cent	per cent	per cent	per cent	
Compact heavy clay.....	86.7	47.6	48.0	0.4	Yes
Clay loam.....	67.2	41.1	46.1	5.0	Yes
Clay loam.....	53.4	33.9	40.7	6.8	Yes
Compact loam.....	46.5	34.9	41.1	6.2	Yes
Very fine sand.....	48.4	39.3	49.3	10.0	No
Friable loam.....	42.6	37.1	49.3	12.2	No
Friable fine sandy loam.....	39.6	34.6	49.5	14.9	No

of this group of soils is somewhat correlated with texture. This fact probably explains why many central European investigators emphasize the importance of particle size in determining the depth and spacing of tile drains. This relationship may or may not always be true, depending upon the structure. Consequently, the character of the pore space is a much more reliable index than texture for judging drainage

needs. These results also point out the fact, which has been repeatedly emphasized in the discussions on soil structure, that the air capacity of soils is not necessarily related to total porosity.

The experiences of various investigators on drainage problems associated with clay soils bear out Kopecky's general ideas on the relation of air capacity to drainage. For example, to cite one specific instance among many observations, two alluvial soils within the same flood plain were found to have entirely different drainage problems even though their texture, consistency and fertility were almost exactly alike. The one produced satisfactory crops with adequate surface drainage; the other did not produce good crops. The former could be drained by lateral ditches with a fairly wide spacing; the latter demanded a very close spacing of such ditches. Porosity measurements revealed that the better-drained soil had an air capacity of 8.1 and 5.9 per cent at depths of 0 to 16 and 16 to 30 inches, respectively. The values for the other soil were 5 and 2 per cent for the same depths.

The experiments of Neal (9) on the drainage of Minnesota soils suggest that the tension-free pore space is an important criterion for estimating the correct depth and spacing of tiles. Although his studies do not specifically mention soil-pore space, the fact that spacing and depth vary inversely with the moisture equivalent implies a relationship with porosity. Neal found that the spacing of tiles changed according to the ratio $k/(\text{Moisture equivalent})^{1.55}$; depth changed according to the relationship, $k/(\text{Moisture equivalent})^{0.5}$. In other words, those soils that hold considerable water in the pore spaces at a pF of about 2.7 are more difficult to drain than those that have larger pores and hold less water at this same tension.

There is reason to believe that much more information on the relation of the air capacity of various soils to drainage will result from continued research in this particular field. One should always keep in mind, however, that the pore-space relationships of the different soil horizons contribute to drainage, and that one particular layer may be the dominant factor in controlling the drainage of the entire soil.

COMPOSITION OF THE SOIL AIR

The composition of the soil air depends primarily upon the extent of biological processes and the ease with which it is renewed by exchange with atmospheric air. Growing plants as well as most soil microorganisms utilize oxygen from the soil air and give off carbon dioxide. It is obvious, therefore, that the soil air should contain more CO_2 and less O_2 than the atmosphere. Moreover, this difference should

be greater the more difficult the renewal of the soil air from the atmosphere.

Russell and Appleyard (12) studied the composition of the soil air under various conditions of cropping and fertilization. They drove a hollow cylindrical tube into the soil and extracted the air by means of a simple mercury pump arrangement. They gave the following percentages by volume as representing the mean composition of soil air: $N_2 = 79.2$, $O_2 = 20.6$ and $CO_2 = 0.25$; for atmospheric air these percentages were: $N_2 = 79.0$, $O_2 = 20.97$ and $CO_2 = 0.03$. The CO_2 content of the soil air is about 6 to 7 times greater than that of the atmosphere. The oxygen content of the soil air is slightly less than that of the atmosphere. These values change with season, soil, crop, tillage and biological activity.

Factors Affecting Composition of Soil Air. SOIL PROPERTIES. The major effects of soil properties are associated with those factors that enhance the air capacity and air permeability of the soil. They are texture, structure, moisture content and the amount of organic matter present.

Wollny (14) showed that the CO_2 content of the soil at $30^\circ C$. increased about 10 times as the moisture content changed from 6.8 to 26.8 per cent. Granular soils contained less than one-half as much CO_2 as powdery soils. Loams contained more CO_2 than sands; the amount of CO_2 increased with depth.

Recent experimental findings of Boynton and Reuther (2) clearly illustrate the effect of soil moisture and depth upon the CO_2 and O_2 composition of the soil air. Typical results, which they obtained by placing a "well" of pyrex tubing in the soil and pumping out the soil gases with a mercury pump, are given in Table 32. This particular soil has a dense subsoil which causes poor underdrainage. It is seen that the CO_2 percentage increases with depth at all seasons of the year. The reverse is true for the O_2 content, since it decreases with depth. Rather wide differences are found between the composition of the air in the surface and the sixth foot. The fact that there were only 0.2 per cent O_2 and 15.5 per cent CO_2 at a depth of 6 feet on November 14, 1937, indicates distinctly anaerobic conditions. It is interesting to follow the O_2 variations below the surface foot in relation to soil moisture. During the wet winter months the O_2 content varies from 0.15 to 0.25 per cent. As the soil dries out after April, the O_2 supply of the lower depths increases rapidly.

The results of Wollny, Boynton, Reuther and others indicate that rather wide variations may be expected in the CO_2 and O_2 content of the air within the pores of various soils, depending upon soil condi-

TABLE 32

O₂ AND CO₂ CONTENT IN THE SOIL AIR AT DIFFERENT DEPTHS IN A SILTY LOAM SOIL (BOYNTON AND REUTHER, 2)

Date	Percentage composition of extracted air											
	1 foot		2 feet		3 feet		4 feet		5 feet		6 feet	
	CO ₂	O ₂	CO ₂	O ₂	CO ₂	O ₂	CO ₂	O ₂	CO ₂	O ₂	CO ₂	O ₂
Nov. 14, 1937.....	1.2	19.4	2.4	11.6	6.6	3.5	9.6	0.7	10.4	2.4	15.5	0.2
Mar. 23, 1938.....	0.15	20.15	2.1	0.15	5.6	0.3
April 21, 1938.....	1.9	18.65	3.2	0.35	6.85	0.25
May 24, 1938.....	3.7	16.2	3.95	13.95	5.6	13.35
June 21, 1938.....	1.7	19.25	4.15	17.1	5.35	16.4	9.6	9.45	10.95	2.1
July 25, 1938.....	2.0	19.8	3.1	19.1	5.2	17.5	9.1	14.5	11.7	12.4	12.6	9.8
Aug. 19, 1938.....	2.4	19.0	3.7	17.4	5.0	16.7	8.55	15.25	11.85	12.95	11.9	11.85
Sept. 22, 1938.....	3.0	15.3	4.8	11.0	5.8	9.95	8.45	8.95	10.55	8.4	10.6	9.0

tions. It is noted, however, that any soil characteristic that is responsible for a large air capacity, such as a coarse texture or granular structure, which is maintained relatively free of water, usually favors a lower CO_2 and a higher O_2 content of the soil air. This relationship is associated with the easy renewal of soil air in soils with large pore spaces.

EFFECT OF CROPPING. The presence of growing plants tends to reduce the O_2 content of the soil air and increase the amount of CO_2 . Ordinary respiration activities of the plant cause an intake of O_2 and an outgo of CO_2 . Möller (8), as early as 1879, showed that the soil air from cropped land contained about 10 times more CO_2 than air from fallow land. Some typical effects of cropping and the addition of manures on the CO_2 content of the soil air are shown by the data of Russell and Appleyard in Table 33. It is noted that there is about 4 to 6 times more CO_2 in wheat land than under fallow. Russell and Appleyard attributed this higher CO_2 content to less favorable exchange with the atmosphere rather than to the CO_2 given off during the respiration of plants. Adding manure to the fallow soil doubled the CO_2 content of the soil air. Since plants require ample amounts of O_2 from the soil air and since CO_2 is given off by the roots during plant growth, it becomes obvious that adequate aeration to permit the replacement of accumulated quantities of CO_2 around the plant roots is essential for plant growth.

ORGANIC MATTER AND BIOLOGICAL ACTIVITIES. Peats usually contain more CO_2 than mineral soils. The data in Table 33 show that

TABLE 33

EFFECT OF CROPPING AND MANURE ON CO_2 CONTENT OF SOIL AIR (RUSSELL AND APPELEYARD, 12)

Treatment	Per cent of CO_2 in soil air at different dates				
	May 15	May 25	June 10	July 7	July 27
Unmanured fallow.....	0.10	0.07	0.08	0.08	0.09
Manured fallow.....	0.22	0.32	0.17	0.36	0.35
Cropped to wheat.....	0.61	0.32	0.35	0.48	0.30

manuring increases the amount of CO_2 in the soil air. In other words, microbiological decomposition is an important factor in CO_2 production. This fact is rather obvious when one considers that CO_2 production is used as a measure of biological activity. The importance of

bacterial activity on the production of CO_2 has long been recognized. Wollny (14) was of the opinion that the amount of free CO_2 in the soil is primarily dependent upon the decomposition of organic substances as modified by physical properties of the soil, such as warmth, moisture and porosity. Russell and Appleyard have attributed the major variations in the CO_2 content of the soil air to microbiological activities.

Romell (11), in discussing the CO_2 content of soils, estimates from Waksman's curve of the effect of soil depth on bacterial numbers that, if gas exchange between the soil air and atmosphere is excluded, the CO_2 concentration at a depth of 20 cm. will double in $1\frac{1}{2}$ hours and increase 10 times in 14 hours. Assuming Sjöström's (11) value of 7 liters of CO_2 produced per square meter per day, there would have to be a complete renewal of the soil air every hour to a depth of 20 cm. in order to maintain its usual average composition. Romell termed this the normal aeration of the soil. He evaluated the various factors affecting aeration on the basis of the extent of this renewal.

It seems sufficiently evident, therefore, that large differences in the CO_2 concentration of the soil air can occur as a result of variations in microbiological activities, especially as soil properties retard or enhance the renewal of the soil air from the atmosphere.

SEASONAL VARIATIONS. The data in Tables 32 and 33 show that there are considerable seasonal fluctuations in the CO_2 and O_2 content of the soil air. These variations are primarily due to temperature and moisture changes. The results of Boynton and Reuther, which are typical of other investigators, indicate that the CO_2 content of the soil air in the surface foot is lower in the colder months and higher in the warmer. The low bacterial activity during the winter causes little CO_2 formation. The low O_2 percentages of the subsurface layers are definitely associated with high moisture contents resulting from low vapor losses during the winter and early spring months. Not until the soil dried out during the summer did the O_2 content of the subsoil begin to increase materially.

Russell and Appleyard studied in detail the causes of fluctuations in the composition of the soil air under English conditions and came to the conclusion that these variations were chiefly due to fluctuations in the rate of biochemical changes. From November to May, temperature seemed to be the determining factor in CO_2 production. From May to November, CO_2 production followed the rainfall; soil moisture was the limiting factor during the warmer months. A maximum in the CO_2 content of the soil air was observed in the late spring and fall; a minimum was found in summer and winter.

RENEWAL OF SOIL AIR—GAS EXCHANGE

Most of the present data on the composition of the soil air indicate that the amounts of CO_2 and O_2 do not vary much within the immediate surface. There may be some question as to the applicability of these data to all soil conditions, especially to rather impervious soils. Nevertheless, in light of the intensity of CO_2 production through plant growth and microbiological activities, the apparent lack of any large accumulation of CO_2 in the surface soil (often true of the subsoil) suggests a rather rapid exchange of gases with the atmosphere. The renewal of the soil air is brought about by diffusion and the meteorological factors: soil temperature changes, barometer variations, action of the wind and changes in the amount of pore space occupied by air as a result of the entrance of rain or irrigation water.

Effect of Meteorological Factors. The influences of the various climatic factors on the renewal of soil air are not continuous; they are intermittent. For example, it is obvious that constant temperature and pressure in the absence of wind or rainfall would not cause any appreciable exchange between the soil air and the atmosphere. Variations in any or all of the factors set up the necessary conditions for gas exchange.

SOIL TEMPERATURE EFFECTS. Temperature may influence the renewal of soil air in two ways. First, there may be temperature differences within the soil between the different layers. It is possible that the contraction and expansion of the air within the pore spaces as well as the tendency for warm air to move upwards may cause some exchange between the various horizons and perhaps with atmosphere. In the second place, the soil and the atmosphere usually have different temperatures. This temperature differential should permit an exchange between the atmosphere and the soil air in the immediate surface.

It is difficult to estimate the significance of temperature effects on gas exchange in soils. Romell (11) has suggested that daily variations of temperature within the soil are responsible for less than $\frac{1}{800}$ of the normal aeration. Temperature differences between soil and atmosphere were considered to be responsible for not more than $\frac{1}{240}$ to $\frac{1}{480}$ of the normal aeration. Thus, it appears that temperature is a minor factor in soil aeration.

BAROMETRIC PRESSURE EFFECTS. Theoretically, according to Boyle's law, any increase in the barometric pressure of the atmosphere should cause a decrease in the volume of the soil air. This diminution in volume should permit an equivalent amount of atmospheric air to

penetrate the soil pores. On the other hand, any decrease in barometric pressure should produce an expansion of the soil air and cause part of it to enter the atmosphere above the soil. Thus, a rinsing of the soil with air should occur from time to time. These statements are predicated upon the fact that any change in barometric pressure of the atmosphere is readily reflected within the soil pores.

Buckingham (3) calculated the possible rinsing action due to barometric changes and showed that penetration of atmospheric air within a permeable soil column 10 feet deep would amount only to about 0.12 to 0.22 inch, dependent upon the magnitude of the barometric change. Thus, it is seen that fluctuations in atmospheric pressure have little influence upon soil aeration, even where there is ready access between the soil air and atmosphere. Romell has estimated that not more than about $\frac{1}{100}$ of the normal aeration of soils can be attributed to variations in barometric pressure.

WIND ACTION. One might expect that the pressure and suction effects of high winds would exert some influence upon the renewal of the soil air. Romell has given some attention to this phase of soil aeration and has come to the conclusion, from calculations based upon wind velocities, that wind action could not be responsible for more than about $\frac{1}{1000}$ of the normal aeration on vegetated soils. It is possible that this value is higher for bare, unprotected soils that are extremely porous. Nevertheless, it is a relatively small part of the total aeration in any case.

RAINFALL EFFECTS. The infiltration of rainfall into the soil may cause a renewal of the soil air in two ways, namely, the displacement of air in the pores by the water, which is subsequently displaced again with air, and the carrying in of dissolved oxygen in the water. One can visualize a rather complete renewal of the soil air following a rain, especially if the water is able to displace the major portion of the air within the pores. In many instances, however, a considerable amount of entrapped air remains, which is not forced out of the soil by infiltrated water. Renewal of the soil air through rainfall effects is periodical, dependent upon the distribution of the rains. Romell has estimated that rainfall accounts for only about one-twelfth to one-sixteenth of the normal aeration.

Effect of Diffusion. Buckingham (3) was one of the first investigators to apply the kinetic theory of the diffusion of gases to soil aeration. Since, according to the kinetic theory of gases, the molecules of gases are in a state of movement in all directions, two gases will readily mix as the molecules of each gas move into the space occupied by the other. This process is usually known as **diffusion**. Inasmuch as the

soil air tends to contain more CO_2 and less O_2 than the atmosphere, the diffusion process in soils consists primarily in the movement of CO_2 out of the soil into the atmosphere and O_2 from the atmosphere into the soil. If this action is complete until equilibrium is attained, there will be a tendency for the soil air to approach the composition of the atmosphere.

Buckingham applied the term **diffusion constant** to designate the rate of flow of gases through the soil-pore space as a result of kinetic movements. This constant was defined as the average number of cubic centimeters of each of two gases which would pass in opposite directions through a layer of soil one centimeter thick, per second per square centimeter of cross section, when the partial pressure of each gas was one millimeter of mercury greater on its respective side of the layer than on the other. Each gas diffuses through the soil layer to the side where its partial pressure is lower.

Buckingham studied the diffusion of air and CO_2 through layers of various soils at different moisture contents. He also measured the transmission of air and CO_2 through these same layers under small pressures. The free pore space was also determined. His results are given in Table 34.

These data show a definite correlation between the free pore space (total pore space not filled with water) and the diffusion constant. The addition of water to the soil layer caused large decreases in the rate of diffusion. These results confirm the findings of Hannen (6), who showed that the sum of the cross-sectional area of the effective pore volume was the most important factor affecting the diffusion of CO_2 . On the basis of these data, Buckingham expressed the relation of the diffusion rate to the free pore space by the equation:

$$D = kS^2 \quad (44)$$

where D is the diffusion constant, S is the free pore space and k a proportionality factor or diffusion coefficient, which was calculated to be 2.16×10^{-4} . The value of the diffusion coefficient varies directly with the square of the temperature (absolute) and inversely with the total pressure. This expression points out that the rate of diffusion is reduced to one-fourth as the free pore space is reduced to one-half.

The interesting feature of this expression is the fact that a free pore space of 1 gives a diffusion constant of 2.16×10^{-4} for the free diffusion of CO_2 and air. This value approximates that of 2.20×10^{-4} found by Loschmidt (3) and that of 2.13×10^{-4} determined by

Obermayer (3). Buckingham realized that it was dangerous to use $S = 1$ in his equation to characterize the free diffusion of these gases. However, the fact that the value thus obtained is so nearly the same as those that were accurately determined suggests that the diffusion

TABLE 34

RELATION OF THE DIFFUSION AND PERMEABILITY OF GASES THROUGH SOILS TO THE EFFECTIVE PORE SPACE (BUCKINGHAM, 3)

Kind and condition of soil	Free pore space, per cent	Weighted diffusion constant $\times 10^5$	Permeability constant
Dune sand—4.8 per cent H_2O	56.1 46.0	8.26 4.71	14.0 2.15
Garden loam—air dry	48.5 43.0	4.73 3.67	0.254 0.101
Garden loam— H_2O added	31.2	2.56	
Garden loam—18.6 per cent H_2O	49.1 39.9 31.9	5.29 3.93 2.64	4.80 1.43 0.445
Cecil clay—air dry	60.9 52.0 46.8	5.26 3.08 2.66	2.20 0.508 0.114
Cecil clay—19.6 per cent H_2O	47.5 35.1 24.9	4.41 2.36 1.80	22.9 5.07 1.25
Cecil clay—20.1 per cent H_2O	46.3 34.7	3.89 1.93	21.9 2.77
Cecil clay—(water added)	25.3	0.61	0.937
Windsor sand, subsoil, 4.2 per cent H_2O	54.9 33.1	7.50 3.28	21.3 1.90
(water added)	16.3	0.51	0.387

process in soils is similar in nature to free diffusion. Apparently, the only controlling soil factor on the rate of diffusion is the free pore space, since the nature, texture, granulation and moisture content of the soil do not materially affect diffusion, only as the free pore space

is altered. This could imply that considerable aeration may take place in heavy clay soils if the free pore space is not too greatly restricted.

The essential difference between the experimental conditions under which the aforementioned results were obtained and the soil is that diffusion of CO_2 and O_2 in the soil air takes place in the presence of N_2 , which makes up nearly 80 per cent of the total gases present. The major effect of the presence of a practically constant quantity of N_2 is to increase the rate of diffusion of CO_2 and O_2 by about 12 per cent. Since the free diffusion of CO_2 and O_2 through N_2 is 7.3×10^{-4} at 25°C . and 760 mm. pressure, it is easy to calculate the diffusion and constant pore space relationship from the expression, $D = 7.3 \times 10^{-4} S^2$.

For a given free pore space content, the volume of CO_2 that diffuses from the soil air to the atmosphere is proportional to the gradient of the CO_2 concentration. This fact is well illustrated by the data of Buckingham in Table 35. It is seen that 10 per cent less CO_2 at a given depth causes a 10 per cent decrease in the amount of CO_2 diffusing to the soil surface. The rate of diffusion, however, is not affected by the concentration gradient.

TABLE 35

RELATION BETWEEN THE VOLUME OF CO_2 DIFFUSING TO THE SOIL SURFACE AND THE EFFECTIVE POROSITY WITH A CO_2 CONCENTRATION GRADIENT OF 0.1 PER CENT PER INCH (BUCKINGHAM, 3)

Effective porosity	0.2	0.3	0.4	0.5	0.6	0.7
Volume of CO_2 diffusing to surface. Cu. ft. per day per sq. ft. (760 mm., 25° C.)	0.0082	0.019	0.033	0.052	0.074	0.10

Unlike the other factors affecting gas exchange between soil air and atmosphere, diffusion is a continuous process as long as the free pore space does not become zero. Changes in the free pore space may result from compaction or saturation of the pores with water. Hansen (6) found that the diffusion of CO_2 through kaolin and a powdery loam was diminished by as much as 40 per cent by compaction. He also observed that moistening sands to 40 per cent of their total water-holding capacity resulted in about a 40 per cent decrease in the rate of diffusion. Diffusion was practically stopped at a saturation of 80 per cent.

Romell (11) has interpreted the data of Hannen and Buckingham as showing that the high water-holding capacity of fine-textured soils restricts the diffusion process considerably under natural conditions. He has suggested that the diffusion coefficient may be lowered from one-half to one-third of that for free diffusion. Moreover, he has emphasized the fact that the presence of a compacted layer at the immediate surface materially restricts the rate of CO_2 diffusion from the lower depths to the atmosphere. In spite of these reductions, however, the amount of exchange resulting from the diffusion process exceeds by far the exchange due to other factors.

PERMEABILITY OF SOIL TO AIR

Since diffusion processes apparently account for most of the normal aeration of soils, one should expect a close relationship between the soil factors affecting the rate of diffusion and the factors that determine the permeability of soil to air. Soils possessing low permeabilities to the movement of air under pressure will undoubtedly not permit rapid diffusion of the soil air to the atmosphere and vice versa.

The importance of air permeability as a physical characteristic of soils was recognized by Renk (10), Ammon (1) and Wollny (17) as early as 1879. These investigators showed that permeability decreased greatly with increasing depth of the soil layer and increasing moisture content; permeability increased with the amount and size of the soil pores.

Porosity and Air Permeability. Any discussion of the relation of porosity to the movement of air through soils must of necessity take into consideration soil-moisture changes, since the effective pore space for air movement within a given soil increases and decreases inversely with moisture fluctuations. Therefore, let us visualize the effects of particle size and arrangement as well as of soil moisture on air permeability, as they relate to the effective soil pore space.

Renk (10) showed that the size of the pore is of utmost importance in the permeability of the soil for air. In comparing soil types of different texture, he found that coarse sand was about 1000 times more permeable to air than fine sand. The total pore volume of the latter was 55.5 per cent, while that of the former was 37.9 per cent. The larger pores in the coarse sand were more effective for air movement. Renk also showed that permeability was greatly lessened by the presence of water in the soil pores. As would be expected, the permeability of those soil types that had a large water-holding capacity was affected most by the presence of moisture.

The findings of Renk were later confirmed by Ammon (1) and Wollny (17). The latter demonstrated that a granular loam was 50 to 100 times more permeable than the same soil in the powdered state (smaller than 0.25 mm. diameter). Wollny also showed that a layer of powdered loam, 1 cm. thick, placed in the middle of a 50-cm. column of quartz sand, lowered the permeability of the entire column 10 to 50 times, depending upon the size of the sand particles. These results emphasize the importance of an impermeable layer near the soil surface in restricting the normal aeration of the entire profile. Buckingham (3) determined the permeability of CO_2 and air through the soil under small pressures and found that the permeability constant was proportional to about the sixth or seventh power of the free pore space. The permeability coefficient, or proportionality factor, was dependent upon the texture, state of granulation and moisture content. It is noted from the data in Table 34 that the addition of water to various soils decreased the permeability constant much more than the diffusion constant.

More recent investigations by Buehrer (4), on the flow of air through columns of lead shot, have shown that permeability varies directly as the square of the average diameter of the particle. In other words, as the size of the pores increases permeability becomes much greater. Experiments on sand-clay mixtures have indicated that only a small amount of clay is necessary to restrict air movement through the system. The effective diameter of the pores between the sand grains is quickly reduced by the clay particles. Buehrer observed that air permeability is not a linear function of porosity. He has suggested that a large portion of the pore spaces between the solid particles does not contribute to the passage of air. The smaller pores apparently offer considerable resistance to air flow.

Most of the experimental evidence seems to suggest that the factors of porosity that limit air permeability are similar to those affecting water movement. The amount and size of the larger pores appear to be of great importance. The presence of water in the soil pores restricts air flow, temporarily at least. Romell (11) has estimated that this restriction may be as much as $\frac{1}{10,000}$ of the normal aeration. The presence of a thin compacted layer in the upper layers may reduce air movement through the profile to an exceedingly slow rate.

Depth of Soil Column and Permeability. The permeability of a given soil column decreases rapidly with the thickness of the layer. Renk observed that the volume of flow through sand columns was inversely proportional to the depth of the column. A similar relationship was obtained by Buehrer. He has shown that this inverse relation

holds more closely with the finer-textured soils. A significant feature of Buehrer's data is found in the large decrease in the rate of air flow through fine-textured soils, after the depth exceeds about 2 to 5 cm. These results, along with the restricted diffusion under compact conditions, suggest that the presence of an impermeable layer at the soil surface may affect soil aeration to a greater extent than is usually realized.

Aeration of the Root Zone of Plants. An analysis of the existing information upon the aeration of soils suggests that many of the conclusions that have been made are not supported by enough substantial facts. Rather good experimental evidence is available to show that surface compaction retards greatly the normal processes of air renewal, as evidenced by decreased permeability to air. On the other hand, CO_2 and O_2 analyses of the removed soil air have usually not indicated any serious concentration of CO_2 or depletion of O_2 . At first glance it seems that these two facts are contradictory. They are, if one uses the composition of the extracted air as an index of aeration. On the other hand, they can be reconciled if analyzed separately.

The rate of air movement from the soil to atmosphere, or vice versa, depends upon the air capacity or content of large pores. The percentage composition of a given amount of air removed from the soil usually does not take into consideration the amount of air that is present in a given volume. Moreover, there is reason to doubt if the composition of the extracted air represents that in contact with plant roots. Therefore, one can hardly call a soil well aerated that contains only 3 per cent by volume of total soil air even though its composition might seem satisfactory.

It is usually assumed that there are sufficient cracks through surface crusts or compacted layers to permit adequate diffusion of air between the soil and the atmosphere. There is little evidence to prove or disprove this assumption. Observations of plant growth on compacted soils throw considerable doubt on the assumption of adequate aeration through this means. Undoubtedly, more research is needed on the plant physiological aspects of soil aeration. Until convincing experimental data are available, the suggestion of Lundegardh that the composition of the extracted soil air is not a true measure of soil aeration seems to carry considerable merit. Moreover, it is believed that the presence of compacted layers restricts aeration to a greater extent than is generally believed.

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CHAPTER VIII

SOIL TEMPERATURE

Soil temperature is one of the more important factors that control microbiological activity and the processes involved in the production of plants. It is a well-established fact that the rate of organic-matter decomposition and the mineralization of organic forms of nitrogen increase with temperature. Consequently, the amount of organic matter that remains in soils is greater under lower than under higher temperatures. Other important microbiological processes also vary in their intensity with temperature; there seem to be certain optimum soil temperature ranges for these processes.

Soil temperature first affects plant growth during the germination of seeds. Seeds of different plants vary in their ability to germinate at low temperatures. Germination is a slow process in a cold soil. Practical experience has shown that germination becomes more rapid as the temperature of the soil rises, up to a certain optimum temperature. The quicker the germination of the seeds, the earlier will be the crop. Thus, cold or warm soils in the spring have different agricultural significance. This is especially true in the vegetable-growing industry.

The growth of the plant is also influenced by soil temperatures. A cold soil is not conducive to the rapid growth of most agricultural plants. Not only are the biological processes that help to provide nutrients for the plant impeded but also the normal growth processes of the plant itself. Naturally, various plants differ in their ability to grow in cold soils. The same is true for conditions of extreme warmth, although the direct effects of too much warmth are usually not as serious as the influence of cold.

Thus, even as Van't Hoff found that the speed of a chemical reaction is doubled by each 10° C. rise in temperature, the life processes within the soil also are accentuated as the temperature rises.

SOURCE AND AMOUNT OF HEAT

The temperature of the soil is primarily dependent upon the amount of radiant energy that is received from the sun. The quantities of heat that reach the earth's surface by conduction from within the earth or

that result from chemical and biological processes are so small that they have a negligible effect upon soil temperature.

The amount of radiation reaching the earth from the sun depends upon the amount emitted by the sun and the absorption of heat rays by the atmosphere. The presence of dust particles and water vapor in the atmosphere decreases the radiation that reaches the earth. Water vapor absorbs large quantities of heat and a moist or cloudy atmosphere prevents much of the sun's radiation from reaching the earth. Likewise, such an atmosphere does not permit a large loss of heat from the earth by radiation. Clear, dry atmospheres are usually responsible for large extremes in temperature; the days are hot and the nights are cool. Moist, cloudy atmospheres are more conducive to uniform temperatures.

In addition to atmospheric conditions which limit the amount of the sun's radiation reaching the earth, there are several characteristics of the earth's surface that greatly affect the amount of radiation that is retained. These may be grouped into (1) those characteristics of position that are related to the angle at which the sun's rays hit the earth, such as the slope, exposition of the land, and latitude, (2) those features associated with the distribution of land and water, as well as water currents, (3) those factors associated with the nature of the soil and the vegetation growing upon it and (4) the effect of elevation or altitude.

Latitude and Slope of Land. The angle at which the sun's rays meet the earth greatly influences the amount of radiation received per unit area. Radiation reaching the earth at an angle is scattered over a wider area than the same radiation striking the earth's surface perpendicularly. Consequently, in the former case the amount of heat received per unit area is decreased in proportion to the increase in area covered. The amount of radiation reaching the earth per unit area is proportional to the cosine of angle made between the perpendicular to the surface and the direction of the received heat rays. Therefore, the radiation received per unit area decreases with an increase in this angle.

The effect of latitude on the angle at which the sun's rays strike the earth is well illustrated in the differences in temperature between the arctic and tropic regions. Even though days and nights may be of equal lengths in both regions at certain periods of the year, latitude has an enormous effect upon the temperature of the soil.

The effects of latitude may be simulated on a small scale within a given latitude by changes in the direction of exposure and the degree of slope of the land. For example, the angle at which the rays of the

sun strike a steep south slope is entirely different from that on a steep north slope. These differences in exposure have great ecological and agricultural significance inasmuch as the temperature of the soil is always higher on southern exposures than on northern. Wollny (13), as early as 1878, studied the effect of exposition on the warmth of soils and found that the south exposures were always several degrees warmer than the north slopes. Because this increased warmth was due entirely to the direct action of the sun's rays, greater temperature variations from night to day were observed on the south exposure. He also found that the temperature differences between exposures were greater, the steeper the slope. Direction of exposure, however, was of greater significance than the degree of slope.

The importance of direction of exposure is often apparent by the difference in the type of vegetation on the north and south slopes of hilly to mountainous terrain. This is especially true in the temperate zone. Inasmuch as the soil remains frozen longer on the north exposures than on the south, the soil has a lower temperature and is more moist after frost has disappeared. Species of vegetation adapted to cooler, more moist climates are generally present on the north exposures, to a more or less limited extent.

Exposure is extremely important in the growing of vegetables and fruit. As far as vegetables and fruits that are not susceptible to frost are concerned, southern exposures permit an earlier planting and harvesting of the crop than other exposures. For certain fruits, however, a warm southern slope may often cause flowering to occur before dangers from frost have passed and the entire crop may be lost by freezing.

Distribution of Land and Water. In general, island climates are more uniform than continental climates. The presence of large bodies of water tends to stabilize the temperature because of the high specific heat of water, which is responsible for the absorption of large amounts of heat. In addition, the atmosphere surrounding these bodies of water is highly saturated with water vapor, which reduces the amount of radiant energy reaching the earth. The effect of the absence of these two factors is well illustrated in the temperature changes within a continent. For example, the central plains region and corn belt of the United States are characterized by hot summers and cold winters. Even the summer nights are warm as a result of radiation from the earth. It should be emphasized, however, that the beneficial effects of water on the equalization of temperature are only associated with rather large masses of water. Little can be achieved in this direction through artificial ponds and lakes. This fact is completely over-

looked in many of the statements associated with water-conservation projects.

The influence of water currents on air and soil temperatures is well known. The Gulf current along the coast of Great Britain and the Japanese current along the shores of northwestern United States are examples of warm currents that materially affect the climate of the land they touch. Every student of geography is acquainted with these facts.

Vegetative Cover. The major effect of vegetation on soil temperature is the insulating qualities of plant cover on temperature fluctuations. Bare soil is unprotected from the direct rays of the sun and becomes very warm during the hottest part of the day. When cold seasons arrive, such an unprotected soil rapidly loses its heat to the atmosphere. On the other hand, a good vegetative cover intercepts a considerable portion of the sun's radiant energy which prevents the soil beneath from becoming as warm as bare soil, during the summer. In winter, the vegetation acts as an insulating blanket that reduces the rate of heat loss from the soil. Consequently, a protected soil is cooler in summer and warmer in winter than one that is bare.

In 1883, Wollny (14) showed that the daily variation in soil temperature at a depth of 10 cm. was 2 to 4 degrees greater under bare soil than under a grass cover. The influence of cover was greater the larger the number of sunny days. He also observed that there was little difference between the effects of different crops if they provided the same amount of shading. Ebermayer (1) studied the effect of forest vegetation and grass on the temperature of the soil and found that bare soils undergo much larger temperature variations than protected soils. Trees were observed to exert greater beneficial effects than grass. A summary of Ebermayer's data is given in Table 36. These results are self-explanatory and point out the relative differences in temperature that may be expected between protected and bare soils.

The investigations of Petit (4) have shown that frost penetration is more rapid and its disappearance slower under bare conditions than under grass or surface mulches. This fact is well illustrated in Figure 57. It is seen that the grass cover decreased the rate and depth of penetration of frost, when compared with bare soil. When thawing occurred, frost disappeared from the protected soils sooner than from the bare, owing to the fact that the latter soils were frozen to a greater depth.

These results have been confirmed by many later investigators. It has been shown that overgrazed pastures freeze deeper than those under controlled grazing because the better growth of grass on the

latter gives more protection to the soil. Likewise, there is ample evidence that soils in forests with a good surface litter freeze only to a

TABLE 36

THE EFFECT OF SURFACE PROTECTION ON SOIL TEMPERATURE
(Ebermayer (1), 5-yr. average at Munich, Germany.)

Season	Soil temperature in °C under different cover			
	8-yr. beech	8-yr. fir	Sod	Bare
Winter.....	1.23	1.30	0.96	0.74
Spring.....	6.14	5.19	6.03	5.55
Summer.....	16.89	16.98	18.11	18.74
Fall.....	10.31	10.31	10.20	9.80
Average (0-60 cm.).....	8.64	8.45	8.83	8.70
Difference between yearly maximum and minimum:				
Surface.....	28.7	25.1	35.6	36.1
0-60 cm.....	21.12	20.20	23.58	24.9

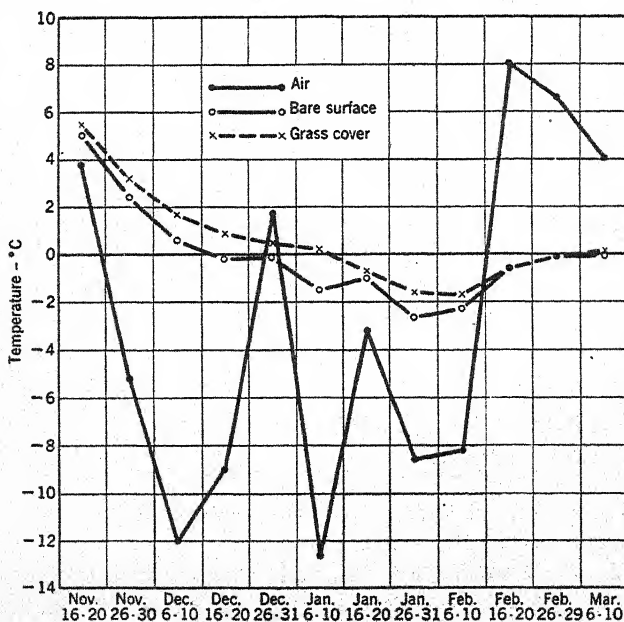


FIG. 57. Frost penetration in bare and grass-covered soils (Petit, 4).

rather shallow depth during the coldest winters. This fact has great hydrological significance, since the soil under good forest conditions is rarely rendered impervious to water by freezing.

Statements have been made which stress the importance of the cooling effects of vegetation, resulting from the transpiration of water from plant leaves, on the temperature of the atmosphere. The original investigations of Wollny (12, 14) aimed to shed some light on this question. He observed that the temperature of the atmosphere 40 cm. above a clover field was slightly lower than that above a fallow field. The differences were not sufficiently large to exert any appreciable direct effect on the soil temperature. These differences were partly due to decreased radiation from the earth under the clover vegetation. It is highly improbable that the transpiration activities of vegetation as such have any significant influence upon soil temperature.

TABLE 37
THE RELATION OF SOIL COLOR TO TEMPERATURE (SCHÜBLER)

Soil type	Natural color	White surface	Black surface
	° F.	° F.	° F.
Yellow-gray quartz sand.....	44.7	43.2	50.9
White-gray quartz sand.....	44.5	43.2	51.1
Yellow clay.....	44.1	42.4	49.7
Loam.....	44.5	42.2	49.5
Blackish-gray humus.....	47.4	42.6	49.4
Blackish-gray garden soil.....	45.2	42.4	50.9

Soil Effects. The temperature of the soil is determined to a considerable extent by its own properties. The temperature of soils depends upon the factors responsible for (1) differences in the intensity of absorption of heat, such as color, (2) variations in the specific heat of the soil, such as composition and water content and (3) differences in heat conductivity, such as compaction and moisture content. Detailed discussions of (2) and (3) will appear in subsequent paragraphs. At this time, attention will be called to differences in temperature that are commonly experienced as a result of the aforementioned factors.

Schübler (15) was the first (1830) to study the influence of color on soil temperature. He covered 4 square inches of various soils with white $MgCO_3$ or black pine soot and determined their temperatures. The results of his investigation are given in Table 37. It is noted that a difference of about 8° C. exists between the white- and black-surfaced

soils. Later, in 1878, Wollny (15) studied the effect of color on soil warmth in greater detail and observed that dark soils are warmer during the warm parts of the year, that the daily variations in temperature are greater with dark soils, that the loss of heat during the night is faster under the darker colors and that temperature differences between dark- and light-colored soils are smaller with increasing depth. He concluded that, although color has a great influence on the warmth of a soil in a dry condition, its effect may often be overbalanced by other properties of the soil. It is important to keep in mind that those properties which are responsible for rapid absorption of heat rays during the day by dark-colored objects also cause rapid emission of heat during the night. This fact explains the daily variations in temperature under the darker colors. Inasmuch as the dark color in most soils

TABLE 38

TEMPERATURE DIFFERENCES BETWEEN VARIOUS SOILS (VON SCHWARZ, 10)

Soil type	Hygroscopicity	Heat capacity of equal volumes			Temperature at the end of 15 min. when soils at 16.5° C. are placed in contact with a heat source at 60° C.	
		Dried at 110° C.	Air-dried	Capillaries saturated with water		
	per cent	cal.	cal.	cal.	dry ° C.	wet ° C.
Peat	21.6	0.140	0.191	0.960	18.7	20.9
Sand	1.4	0.325	0.347	0.675	35.3	56.2
Loam	3.7	0.326	0.341	0.762	32.2	49.3
Clay	9.2	0.289	0.406	0.804	28.0	39.2

is usually due to organic matter, the temperature of these soils is greatly influenced by the relatively high specific heat and low conductivity of organic materials.

The significance of the specific heat and heat conductivity of various soils is shown by the data of Von Schwarz (10) in Table 38. These results show that peat soils are slow to warm up because of a low heat conductivity, due to the large amount of air incorporated within the pores. The sand warms up most rapidly because of its greater conductivity. In all cases, the presence of water causes the larger temperature rise, owing to its increasing the thermal conductivity. The materials with the largest specific heat in the moist state exhibit the smallest temperature rise. Petit (4) found that frost penetration was fastest in sand, then in clay and slowest in peat. These data call attention to the fact that specific heat and thermal conduction are properties

that vary considerably with different soils and soil conditions and, therefore, greatly influence soil temperature.

HEAT CAPACITY OF SOILS

The specific heat of any substance is defined as the calories of heat required to raise one gram one degree on the centigrade scale. The heat capacity of a given material is equal to its specific heat times its mass. The specific heat of water is 1.0. All other constituents of soils have much lower specific heats.

The contributions of Lang (2), in 1878, and Ulrich (9), in 1894, on the specific heat of various soil constituents undoubtedly have been the most outstanding of all investigations. Their values are widely used and quoted by numerous workers in the field of soils. It is seen from the data in Table 39 that quartz has the lowest specific heat of the major soil constituents and humus the highest, excepting water. The

TABLE 39
THE SPECIFIC HEAT OF VARIOUS SOIL CONSTITUENTS

Material	Lang (2)	Ulrich (9)
Coarse quartz sand.....	0.198 (0.517*)	0.191 (0.311*)
Fine quartz sand.....	0.194	0.192
Quartz powder.....	0.209	0.189
Kaolin.....	0.233 (0.576*)	0.224 (0.233*)
Calcium carbonate.....	0.206	0.208
Magnesium carbonate.....	0.260	0.246
Humus.....	0.477 (0.601*)	0.443 (0.165*)
Gypsum.....	0.195
Fe ₂ O ₃	0.163	0.165
Al ₂ O ₃	0.217
Fe(OH) ₃	0.226
Orthoclase.....	0.194
Oligoclase.....	0.205
Potash mica.....	0.208
Magnesia mica.....	0.206
Hornblende.....	0.195
Apatite.....	0.183
Dolomite.....	0.222
Talc.....	0.209
Calcareous sandy soil.....	0.249
Humus calcareous sandy soil.....	0.257
Garden soil.....	0.267

* Specific heat according to volume.

aluminosilicate, kaolin, has a slightly higher specific heat than quartz. Since the major constituents in most soils are quartz, aluminosilicates, water and humus, it is evident that humus and water will affect the heat capacity considerably. It is significant to note that the specific heats of the various solid soil constituents, on the volume basis, do not differ greatly.

Lang showed that the heat capacity and specific heat of a soil could be calculated by adding together the capacities of the various constituents. For example, the calculated specific heat of the garden soil in Table 39 was 0.262, as compared with the measured value of 0.267. Therefore, the heat capacity of a soil can be obtained from the expression

$$C = c_1M_1 + c_2M_2 + c_3M_3 + \dots \quad (45)$$

where c_1, c_2, c_3 , etc., are the specific heats of different soil constituents of mass M_1, M_2, M_3 , etc.

The importance of water as a contributing factor to the heat capacity of a soil is clearly shown by the results of Ulrich, when they were expressed on the volume basis; that is, when a given volume of dry soil was saturated with water much the same as would occur under natural conditions. He found that the heat capacities, on the volume basis, of kaolin when dry and when 50 and 100 per cent saturated with water were 0.233, 0.539 and 0.846, respectively. The corresponding values for humus were 0.165, 0.555 and 0.945. Thus, it is seen that wet soils have much higher heat capacities than dry ones. This explains the cold nature of wet soils and emphasizes the need for adequate water removal from soils in order to cause them to warm up earlier in the spring.

HEAT CONDUCTANCE AND FLOW IN SOILS

Since the sun is the source of heat that controls the temperature of soils, it is obvious that the flow of heat in soils is an important factor in determining soil temperature. Heat flow in a porous material such as the soil is much more complicated than flow through solid bodies. Nevertheless, the same fundamental features of heat flow operate in both types of materials.

The Theory of Heat Flow. In discussing the flow of heat through a given substance, let us follow the general ideas used by Patten (3). Consider the case where a metal bar, 100 cm. long, is placed in contact with a heat source so that under conditions of steady flow of heat the one end is at a temperature of 100° C. and the other at 0° C. (See

curve 1, Figure 58.) In order to calculate the amount of heat that flows through any given part of this bar, visualize a small rectangular section within the bar at a distance of x cm. from the heated end. Let the cross-sectional area of this section be A and the thickness of an infinitesimal distance Δx .

If θ is the temperature of the face of this section at x , the flow of heat per unit time through the surface A will be equal to $-KA(d\theta/dx)$, where $d\theta/dx$ is the temperature gradient, or the change in θ with distance from the heat source, and K is the heat conductivity of the material. The negative sign indicates that the temperature decreases as the distance x increases. This expression simply states that the flow of heat from the hot to cold regions of a bar increases directly with the conductivity of the material, the cross-

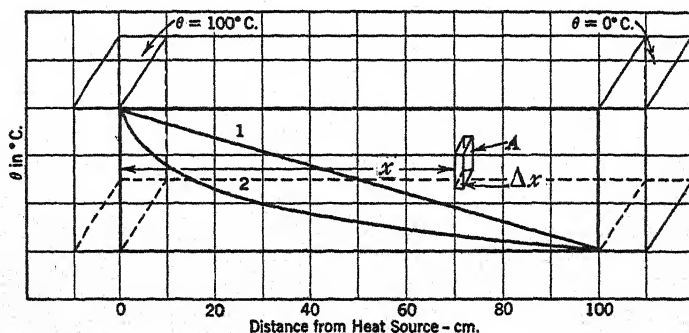


FIG. 58. Sketch illustrating the flow of heat through soils.

sectional area through which heat flows, and the greater the difference in temperature between the hot and cold ends. Heat conductivity is generally defined as the quantity of heat that flows through a unit area of unit thickness in unit time under a unit temperature gradient.

The temperature of the surface of this rectangular section at a distance $x + \Delta x$ from the source of heat will be lower than at x and will be equal to $\theta - (d\theta/dx)\Delta x$. Since $d\theta/dx$ is the rate at which the temperature decreases as the distance, x , from the source of heat increases, then this rate of decrease times the distance, Δx , over which this fall in temperature occurs will give the total drop in temperature in passing from x to $x + \Delta x$. In this case Δx is so infinitesimally small that the rate of temperature fall may be considered constant over this distance. Therefore, the flow of heat through the surface of the section at $x + \Delta x$ will be equal to $-KA[d/dx(\theta - (d\theta/dx)\Delta x)]$. The expression within the outer brackets is simply the rate of change of the temperature at $x + \Delta x$.

Since the temperature at $x + \Delta x$ is lower than at x , there will be less heat leaving the rectangular section than enters. This difference is given by the expression, $-KA(d\theta/dx) - \{-KA[d/dx(\theta - (d\theta/dx)\Delta x)]\}$, which in simplification reduces to:

$$-KA \frac{d^2\theta}{dx^2} \Delta x \quad (46)$$

The term $d^2\theta/dx^2$ is the rate of change of the temperature gradient, or the acceleration of the temperature change with distance. Under a constant gradient in a steady state this acceleration is zero and there is no difference between the amounts of heat entering and leaving the section.

Consider now that Figure 58 represents a box of soil. Before reaching the steady state, the temperature at each point in the soil is changing according to curve 2. Under this condition the heat that flows into one side of the rectangular section is not the same as that which leaves it on the other side. Let the mean temperature of the section rise by a small amount, $d\theta$, in a small unit of time, dt . The quantity of heat that is necessary per unit time to raise the temperature of this section will be equal to $A c (d\theta/dt) \Delta x$, where c is the heat capacity of the soil (c = effective specific heat of soil/apparent specific volume). This expression states that the amount of heat required is equal to the volume of the section, $A\Delta x$, times its heat capacity, c , times the change in temperature, $d\theta/dt$.

If the soil is thoroughly insulated so that there is no heat loss during the experiment, then

$$AK \left(\frac{d^2\theta}{dx^2} \right) \Delta x = Ac \left(\frac{d\theta}{dt} \right) \Delta x \quad (47)$$

or

$$\frac{K}{c} \left(\frac{d^2\theta}{dx^2} \right) = \frac{d\theta}{dt} \quad (48)$$

This is the fundamental equation for calculating the heat conductivity, K , from data on the effective heat capacity, c , the rate of change of the temperature gradient, $d^2\theta/dx^2$, and the change in temperature at a given point with time, $d\theta/dt$.

The expression K/c has been termed **diffusivity** by Patten and denotes the temperature change that takes place in any portion of the soil as the heat flows into it from an adjacent layer.

The theory of heat conductivity may be visualized somewhat more simply by considering the rectangular section in Figure 58 from a

slightly different point of view. Let the temperature on each side of this section be equal to T_1 and T_2 , the thickness of the section equal to d , and the amount of heat that flows across in a given time equal to Q . The rate of heat flow per unit area will be Q/At . The temperature gradient will be $(T_1 - T_2)/d$. Therefore, according to definition, the heat conductivity of a substance will be given by the expression:

$$K = \frac{Q}{At} \bigg/ \frac{T_1 - T_2}{d} \quad \text{or} \quad \frac{Qd}{At(T_1 - T_2)} \quad (49)$$

Soil Factors Affecting Heat Conductance. SOIL COMPOSITION AND POROSITY. Pott (5) was one of the first to investigate the soil differences that affect heat movement. He put the soil in a rectangular, insulated container, one end of which was in direct contact with the source of heat. Six thermometers were placed at intervals of 6 cm. away from the heat source. The relative conductivity (undoubtedly diffusivity with this technique) was determined by the average temperature change over the entire column, as well as by the reading of the last thermometer at the end of 12 hours. These values are given in Table 40 and point out that quartz transmits heat faster than humus or clay. The rate of temperature change throughout the soil column was increased by compacting the various materials and by adding water.

Von Schwarz (10) obtained similar data, which indicated that the heat conductivity of different soils followed the order of sand > loam > clay > peat. The same order of conductivity was observed by Wagner (11). Water increased the conductivity. (See Table 40.) The results of Patten (3) show similar relationships.

More recent investigations by Smith and Byers (8), using rather refined techniques, have confirmed the earlier data relative to the heat conductance of sands, clays and peats. They have shown that the thermal conductivity of the soil constituents, with the possible exception of organic matter, varies little from one soil to another. This is an important finding since it shifts the older emphasis on composition to a recognition of the significance of porosity. The degree of packing and the porosity of the soil seem to be the major factors determining the thermal transfer. Smith and Byers arrived at the following approximate expression for the heat conductivity of a dry soil:

$$k = k_2P + k_1(1 - P) \quad (50)$$

where k_2 and k_1 are the conductivities of dry air and of the soil material, respectively, and P is the porosity. Patten's data on dry soils show this same relationship when they are recalculated on a total

porosity basis. This expression shows that the conductivity of the soil decreases as the porosity increases. This is due to the low conductivity of the increased amounts of air between the solid particles as well as poorer contact per unit volume between the soil materials.

TABLE 40
RELATIVE HEAT CONDUCTIVITY

Material	Relative heat conductivity			Worker
	Air-dry compact	Air-dry loose	Moist loose	
Quartz powder.....	106.7	100	201.7	Pott (5) Quartz powder air-dry, loose = 100
Fine quartz sand.....	104.6	197.9	
Coarse quartz sand.....	110.2	
Kaolin clay.....	96.4	90.7	155.6	
Humus.....	98.1	90.7	94.3	
Kaolin + quartz pebbles.....	109.0	
Kaolin + CaCO ₃ pebbles.....	115.3	Von Schwarz (10) Sand, moist loose = 100
Peat.....	51.5	58.8	
Sand.....	85.5	100	
Loam.....	83.3	99.3	
Kaolin clay.....	74.3	90.3	
Peat.....	67.9	Wagner (11) Quartz, air dry compact = 100
Kaolin clay.....	74.7	
Quartz.....	100	
Loam.....	86.7	143.4 (compact)	

INFLUENCE OF SOIL MOISTURE. In the earlier investigations of Pott, von Schwarz and Wagner, it was observed that the heat conductivity of soils and soil materials increased with the moisture content. Since the technique of measuring heat transference consisted in determining the temperature rise of the soil at a given time, it might be more correct to say that diffusivity (K/c) increased with moisture.

In 1909, Patten (3), of the United States Bureau of Soils, carried out his well-known experiments upon the effect of water on heat conductivity and diffusivity. He used about the same procedure as Pott and Wagner for measuring the temperature changes, throughout a soil column in contact with a heat source, as a function of time. The soil with a known content of water was packed into a well-insulated,

rectangular box by a specially devised packing technique so that the volume weight could be measured and, consequently, the effective specific heat could be calculated. Thermometers were embedded in the soil at 1-cm. intervals from the heat source. The temperature was read at frequent intervals on all the thermometers.

Patten realized that one of the greatest difficulties attached to this method of measurement consisted in the flow of water away from the heat source concurrently with the flow of heat. Since an increase in temperature causes an increase in vapor pressure of the water and a decrease in surface tension and viscosity, soil-moisture equilibria are disturbed near the heat source. Water tends to move to regions of

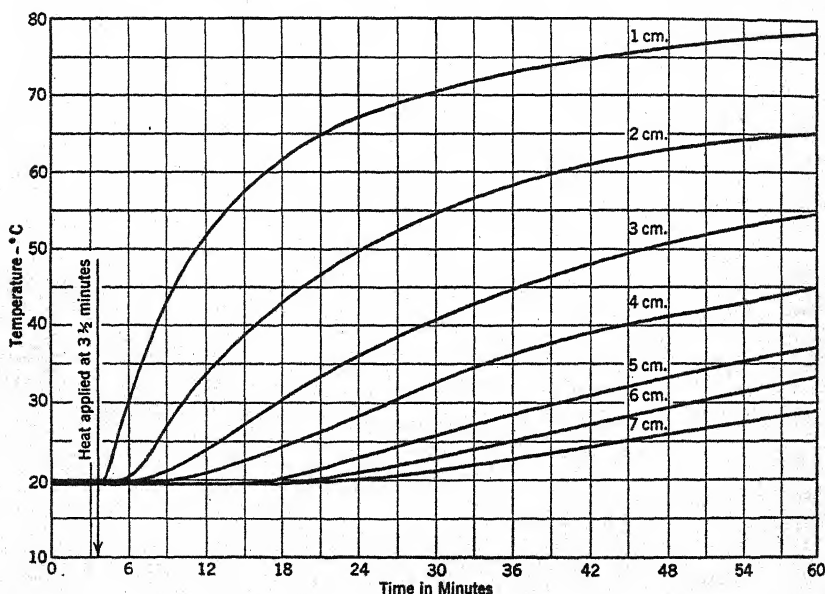


FIG. 59. Temperature changes in soils at different 1-cm. intervals from a heat source as a function of time (Patten, 3).

lower vapor pressure and higher surface tension. This movement of water affects the transfer of heat. In light of this fact, Patten took the readings on the thermometer at 5 cm. from the source at the end of 15 minutes as being somewhat free from the errors caused by moisture flow. They were considered sufficiently accurate to measure the effects of heat transfer.

In order to determine the rate of temperature rise, $d\theta/dt$, at a given distance, x (5 cm.), from the source of heat at time, t , the thermometer readings were plotted as a function of time (Figure 59) on a

large-scale graph. The slope of the curve for the 5-cm. distance was measured by drawing a tangent to the graph at the 15-minute time interval. In order to determine the rate of fall of temperature with changing distance, $d^2\Theta/dx^2$, from the source of heat, the temperature, Θ , of each of the thermometers at the end of 15 minutes was plotted as a function of the distance, x , from the source of heat. Tangents to the curve were drawn at 4.5 cm. and 5.5 cm. to obtain the rate of change of temperature, $d\Theta/dx$, at these distances. The difference between the two rates was then divided by the distance interval, 1 cm., to obtain $d^2\Theta/dx^2$ at 5 cm. Inasmuch as the effective specific heat, c , could be calculated from the respective heat capacities of the water and soil in a given volume, the only unknown in equation 48 was K , the heat conductivity.

Patten studied the conductivity of various soils at different moisture contents. He found that the conductivity of dry quartz particles, as well as soil, was only about one-half to one-third that of water (0.005 cal. per sq. cm. per sec. per degree change in temperature gradient) and about one-fifteenth to one-twentieth that of a solid quartz block. Thus, the conductivity of quartz, for example, is greatly decreased when it is divided into particles; this is due to the small amount of surface contact between particles through which heat will readily flow. The presence of water between the particles increased the conductivity to values higher than that of pure water. The presence of a water film at the points of contact of the particles improved the thermal contact between them and displaced air, a poor heat conductor, with water, which is a much better conductor.

A typical example of Patten's curves is given in Figure 60. It is seen that heat conductivity increases with moisture content. It is also to be noted that the greatest rate of increase in conductivity takes place at the lower moisture contents. On the other hand, with fine quartz powder and fine-textured soils, Patten's original curves indicate that the greatest rate of change of heat conductivity apparently occurred at the higher moisture contents.

Recent experimental results (6) on the use of heat flow as a measure of soil moisture indicate that all soils show the same tendencies with respect to the change in heat conductivity with moisture (Figure 55, page 242). In other words, the increase in heat conductivity is most rapid at the lower moisture contents. At first glance there may seem to be a discrepancy between the results of these two investigations. Essentially, this is not true. The data in Figure 55 were obtained by measuring the change in heat flow as a function of varying moisture at a constant volume weight. That is, the air in a given

volume of soil was displaced with varying quantities of water. Compaction was not materially affected.

Patten determined heat conductivity as a function of changing moisture content in a system that also varied with respect to packing. There was a new packing for each moisture content. This is illustrated in the total porosity curve in Figure 60. Other samples exhibited greater changes in volume with moisture than the one illustrated. He always obtained better packing with dry and extremely wet soils. In

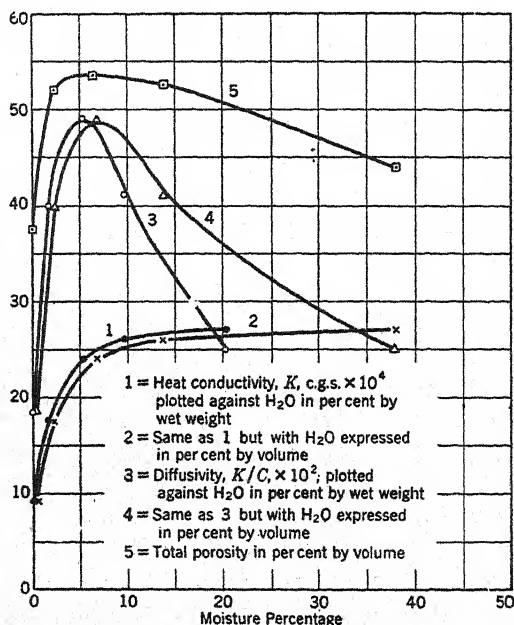


FIG. 60. Heat conductivity and diffusivity in coarse quartz sand (Patten, 3).

the case of quartz sand, where the smallest change in volume occurred with changing moisture content, the heat-conductivity curve was of a different nature from the curve of most of the other samples. Patten recognized that the change in volume weight would affect the results and, consequently, calculated the effective specific heat as well as the conductivity on the volume basis. However, he plotted conductivity as a function of the percentage of water on a wet weight basis. If his data are plotted as a function of the percentage of water by volume, the same general relationship is obtained between the various soils. The curves in Figure 60 illustrate the differences that are obtained

when heat conductivity and diffusivity are plotted as functions of the water content on the wet weight and volume basis.

It is extremely important to recognize the effect of increased contacts, as a result of compaction, on heat conductivity. Wagner (11) observed that the relative difference in heat conduction by a calcareous sandy soil at 16.61 and 22.75 per cent was 245 for a very compact soil and 153 for slight compaction, if the difference with loose soil was taken as 100. Theoretically, if one visualizes two quartz spheres in contact with each other in the dry state, it is seen that the major conduction of heat takes place through a relatively small cross-sectional area at the point of contact. With the addition of a small amount of water in the wedges at the point of contact, the surface through which heat is conducted increases greatly; the distance of flow through the water wedge is small. With further additions of water, the distance of flow through water (the volume of water) increases faster than the augmentation of surface. Therefore, the rate of heat conduction should increase more slowly. This is apparently the case when water gradually displaces the air in the pores of a constant volume of soil.

Returning to Figure 60, we see that the diffusivity, K/c , increases rapidly at first with increasing moisture to a maximum and then decreases. This is due primarily to the greater rise in conductivity at the lower moisture contents as compared with the increase in heat capacity. As the moisture content becomes larger, however, the value of c becomes greater and diffusivity decreases.

DAILY AND SEASONAL CHANGES IN SOIL TEMPERATURE

The movement of heat into or out of the soil depends upon the difference in temperature between the surface of the soil and the lower soil layers. Heat always flows in the direction of the fall in temperature, that is, from warmer to cooler layers. The flow of heat within a soil for a given soil condition will be greater the larger the difference in temperature between any two adjoining layers. Heat will enter the soil as long as a temperature gradient downward exists between the surface and the subsurface layers. Heat will leave the soil as long as there is a temperature gradient upward.

Since the sun is the source of practically all the heat that enters the soil, there will always be a movement of heat downward during the time that the sun's rays strike the soil. If the surface of the soil becomes cooler than the subsurface layers during the night, there will be a movement of heat toward the surface and into the atmosphere,

that is, if the atmosphere is cooler than the soil. Thus, there are a daily rise and a daily fall in soil temperature. The magnitude of this rise and this fall is larger in the summer than in the winter because of the greater heating effects of the more direct sun's rays throughout the warmer months.

The diurnal change in the amount and direction of heat flow is appreciably affected by the nature of the soil and the type of surface protection that is provided by vegetation or artificial mulches. The diurnal changes that take place on protected and bare soils are illustrated in Figure 61. These data of Wollny (14) are fairly typical of

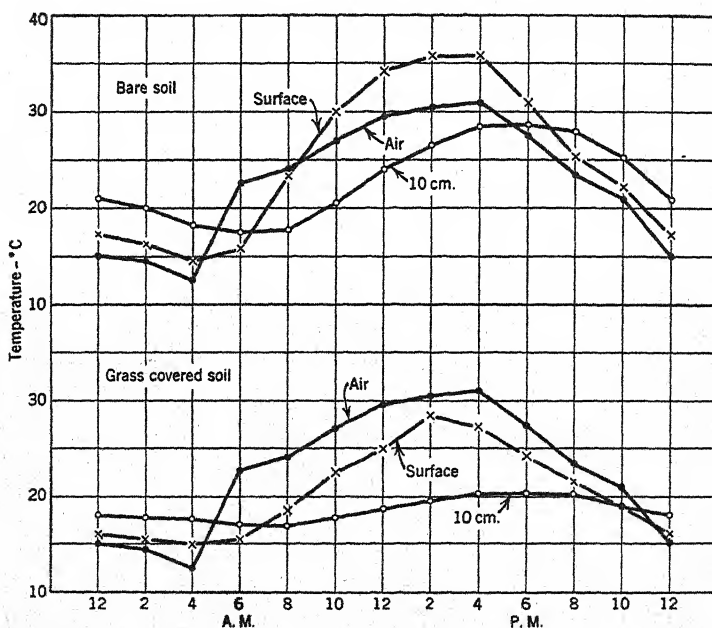


FIG. 61. Diurnal changes in soil temperature in bare and grass-covered soils (Wollny, 13).

the results of other investigators. It is seen that the soil warms up during the sunny portion of the day and cools off at night. From about 6 A.M. to 6 P.M., depending upon the surface protection, heat moves from the surface downward. During the night the reverse process occurs.

There is considerable difference between the daily temperature curves for bare soil and for soil that is covered with grass. With both soils, the atmosphere becomes warmer than the surface at about 5 A.M. From this time until about midnight the two soils vary considerably.

The direct effect of the sun's rays soon raises the temperature of the soil to a point higher than that of the atmosphere. This higher temperature differential of the surface is present throughout the day and night, with the exception of a short time in the early morning (5 A.M. to 8 A.M.). As long as the temperature of the soil is higher than that of the atmosphere, there is a flow of heat from the surface of the soil into the air. The grass-covered soil, on the other hand, never becomes as warm as the atmosphere at any time during the day. The vegetation intercepts the sun's rays and reflects them back into the atmosphere. Heat leaves the protected soil only between about 11 P.M. and 5 A.M.

It is interesting to compare the differences in temperature between the surface and at a depth of 10 cm., in the two soils. In the bare soil, the surface becomes warmer than the subsurface at about 6 A.M. and maintains this higher temperature until about 7 P.M. During this time heat flows from the surface downward; between 7 P.M. and 6 A.M. the direction of heat flow is upward. The maximum temperature at 10 cm. is reached about 2 hours after the maximum in the surface. With the grass-covered soil, the surface is warmer than the subsurface from about 7 A.M. to 10 P.M. Thus, there is an upward flow of heat from 10 P.M. to 7 A.M.

Not only does the grass cover change the time of the occurrence of the diurnal heat waves into and out of the soil but it also affects the magnitude of the temperature changes. This is evidenced from the fact that the major fluctuation of the temperature of the subsurface of the grass-covered soil is only about $3.5^{\circ}\text{C}.$; that of the bare soil is about $11^{\circ}\text{C}.$ Cover tends to equalize the temperature of the subsoil. This is even truer under good forest vegetation than under grass.

In addition to regular daily changes in temperature, the soil undergoes seasonal variations. Here again there is a wave effect of considerably greater magnitude than the daily fluctuations. It is obvious that the soil receives warmth during the summer and loses it during the winter. The extent of the difference between heat gains in the summer and losses during the winter is affected by surface cover. The presence of vegetation, leaf litter or snow greatly influences the amount of heat movement between soil and atmosphere.

The seasonal wave effect of soil temperature is shown by the data of Smith (7) in Figure 62. Under California conditions of sunshine, the soil is warmer than the atmosphere throughout the year. Consequently, there is always heat movement from the soil surface into the air. An upward movement of heat from a depth of 3 feet to the surface takes place from October to March, inclusive; then heat moves downward.

The yearly changes in temperature with depth is more clearly portrayed by the results of Smith (7) in Figure 63. The data for only six months of the year are plotted, for the sake of clarity. Although the magnitude of the temperature variations decreases with depth, there is a slight influence of heat movement downward to the 12-foot depth. Smith found that the temperature of the surface soil reached a maximum in July and a minimum in January. After August the upper layers began to get cooler as the air temperature dropped. However, the lower depths continued to warm up, as is evidenced by the temperature at 12 feet during November. In other words, there was heat movement upward and downward in the profile in the early fall. After November, the flow of heat from the 12-foot depth was upward.

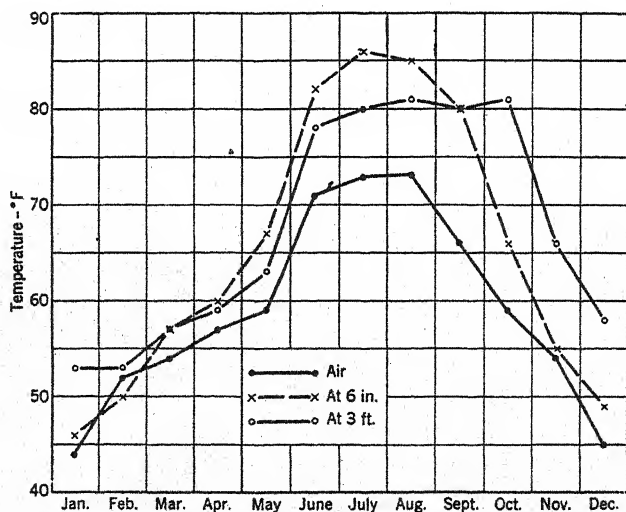


FIG. 62. Seasonal variations in soil temperature (Smith, 7).

Cooling of the entire profile continued throughout January. Then the surface began to warm up slightly. The lower depths continued to cool, as shown by the temperature at 12 feet in March. After March the direction of heat flow was downward.

These seasonal differences in the amount and direction of heat flow in soils are associated with the changes in specific heat and thermal conductivity of the soil profile as a function of moisture contents and temperature gradients. These factors have been discussed in conjunction with the theory of heat flow in soils.

Aside from the more or less regular daily and seasonal fluctuations in temperature, there are other intermittent effects on soil temperature

that are brought about by wind, rain and snow. Direct wind effects on the temperature of the soil are small. However, the evaporation of moisture from the soil as a result of a dry wind will tend to cool the surface, owing to the heat required in the vaporization process.

In general, rain usually exerts a cooling action on the soil because most rains are of a lower temperature than the soil itself. The penetration of cool percolating waters through the soil column may cause rapid cooling effects in the subsoil layers. The occurrence of a warm rain during the winter months will raise temporarily the temperature of the upper layers. Excessive amounts of cool rains during the early

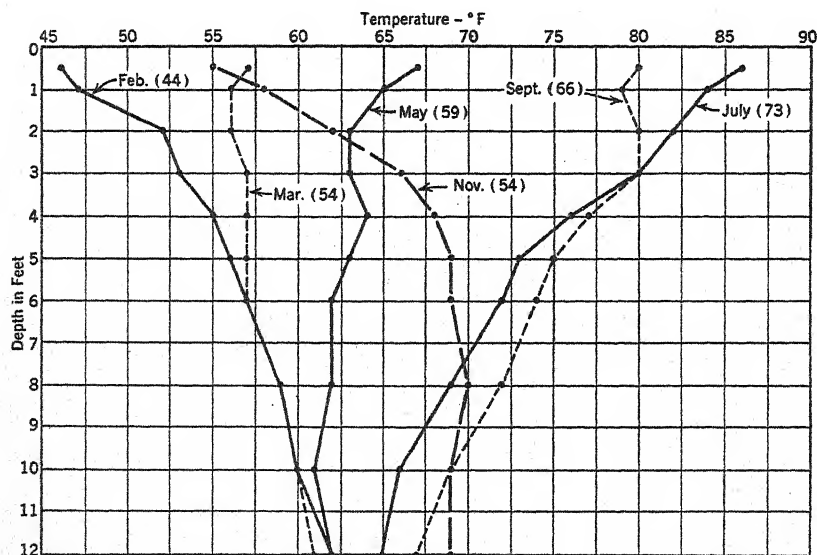


FIG. 63. Seasonal variations in soil temperature at various soil depths (Smith, 7). (Values in parentheses refer to mean monthly temperatures.)

spring months often hinder the warming up of the soil, owing to the high specific heat of water. Most of the effects of rain on soil temperature, however, may be looked upon as intermittent.

A cover of snow on the soil during the winter acts as an efficient insulator to rapid and extensive temperature changes. Farmers are fully aware of the beneficial effects of snow on winter small grains and winter cover crops. All experimental data on the effect of snow cover on soil temperature and the penetration of frost show relatively shallow freezing under snow, unless the air temperature sinks very low for long periods.

Seasonal changes in temperature within the temperate region may be summarized as follows:

	WINTER	SPRING	SUMMER	FALL
Radiant energy from the sun	Low, owing to large angle of contact with earth	Medium, depending upon cloudiness	High, owing to small angle of contact with earth	Medium
Specific heat of soils	High, owing to large moisture content	Medium, depending upon rainfall and evaporation	Low, owing to small moisture content	Medium, depending upon rainfall and evaporation
Conductivity of soils	Good	Good	Fair, dependent upon the extent of drying and loosening of the surface	Good
Characteristics of soil temperature	Surface soil cooler than subsoil	Surface soil slightly warmer than subsoil	Surface soil warmer than subsoil	Surface soil slightly cooler than subsoil
Direction of heat flow	Subsoil to surface to atmosphere	From immediate surface and lower subsoil to upper subsoil	Surface to subsoil	Upper subsoil to lower subsoil and surface

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CHAPTER IX

PHYSICAL PROPERTIES OF SOILS AND TILLAGE

Tillage refers to the different mechanical manipulations of the soil that are used to provide the necessary soil conditions favorable to the growth of crops. It is one of the oldest of arts associated with farming. The evolution of tillage implements and machinery has been one of the most striking features in the growth of the art of farming. Even though phenomenal advances have been made in the perfection of tillage tools, tillage operations remain the most costly of all the various items that are concerned in crop production. Thirty per cent of the total power consumption in agriculture in the United States is expended in tillage (26). Over half this power is used in the basic operations of plowing and listing.

Until rather recently, the problem of tillage had not been approached from the point of view of soil mechanics. Tillage operations were performed without much knowledge concerning the effect of these manipulations upon soil and plant and without any clear picture of the objectives desired. Practical experience was the main guide to the time, kind and extent of tillage sought. One of the best-known earlier assumptions as to the reason for tillage was made by Jethro Tull, an English farmer, in the eighteenth century. He suggested that tillage improved the productiveness of soil because it caused a breaking down of the large soil particles into smaller ones which increased the surface from which plant roots obtained their food. Tull's ideas were responsible for his introducing "horse-hoeing husbandry," which consisted of planting crops in rows and "horse-hoeing" in between.

The modern approach to tillage problems has many far-reaching possibilities. At the moment, the solution of the problem is in only the preliminary stages. As Slipper (21) has suggested, the problem of soil manipulation is of a biologic and dynamic nature. Biologically, it requires fundamental knowledge concerning the requirements of different plants on different soils relative to the physical condition of the soil that is most favorable for plant growth. From the dynamic point of view, tillage operations must be analyzed as to the various physical

processes that take place within the soil, in regard to their effects on the nature of the tillage operation as well as the influence of certain tillage practices upon the activity of certain of these processes.

In previous chapters we have discussed the physical properties of various soils as they relate to certain requirements for plant growth. In the present chapter we will consider the different soil factors that are concerned in tillage operations, as well as the effect of tillage upon certain physical changes of the soil and their relation to crop production.

SOIL TILTH

Tilth is commonly defined as the physical condition of the soil in its relation to plant growth. This definition includes all those soil conditions that determine the qualities of the soil as a suitable physical environment for plant growth. It is obvious that soil-structure relationships constitute a major portion of this physical condition. Adequate aeration, sufficient moisture, ready infiltration of rainfall and so forth are functions of good tilth. But, does tilth refer only to structural relationships? When one analyzes tilth thoroughly he is immediately impressed with the fact that tilth includes also certain phases of soil consistency.

A soil in good tilth from the point of view of structure is generally described in terms of consistency. For example, such a soil is said to be mellow and friable; it handles easily. Friability is one of the major forms of soil consistency. It represents that stage in the soil-moisture curve where there is sufficient moisture around the particles to overcome the cohesive forces between the particle surfaces but not enough of a moisture film to cause plasticity. Therefore, it seems that tilth is defined on the basis of certain structural properties of the soil that are manifested within a given range of soil consistency. In other words, one designates first the consistency range and then defines the soil-structure relationships that are involved in tilth within this range. For example, if a silty clay loam soil is examined when it is abnormally wet or abnormally dry, the impression of good tilth is rarely obtained. The soil does not handle well in either extreme, and conditions are not favorable for good plant growth. Therefore, it is essential to keep in mind that true tilth refers to certain structural conditions of the soil within an optimum range of consistency.

It is significant to point out that the range of friability in which tilth reaches an optimum from the standpoint of plant growth also is the same range in which the soil can be tilled with the least output of power, and with the best effects upon granulation. In other words,

even though tilth is defined primarily on the basis of plant-growth effects, the physical conditions for plant growth so characterized in good tilth are usually, not always, favorable to good tillage. In many instances, tillage is responsible for the achievement of good tilth. These relationships will be discussed later.

Tilth is a dynamic, not a static soil condition. The tilth of a given soil tends to deteriorate under the usual cropping and tillage operations. Good tilth may be renewed through the use of sod crops in the rotation and suitable tillage methods.

Measurements of Tilth. Various methods have been proposed to measure soil tilth. Keen and his associates (9) have attempted to use the size distribution of various clods as a measure of tilth. Soils were screened dry in the field. This technique has been modified recently along the lines of the Russian investigators (11) who wet-sieve the soil in benzene. The English workers attach considerable importance to the amount of material finer than $\frac{5}{8}$ inch in diameter.

Several Russian investigators (11, 22) have emphasized the importance of granulation and porosity as measures of tilth. Most of their results indicate that a seedbed constituted of aggregates 2 to 3 mm. in diameter is the best for plant growth. The soil's pore space should be about equally divided between capillary and non-capillary pores. When the non-capillary porosity is lower than 10 per cent by volume of the total soil, there is poor tilth. A considerable amount of data has been obtained to show that the non-capillary porosity of the soil is a reliable index of tilth in relation to plant growth. Yoder (26) has shown that cotton yields on artificially prepared seedbeds are definitely associated with the non-capillary porosity of these seedbeds. Recent experiments in Ohio have shown that the production of greenhouse soils and sugar-beet soils is closely related to the non-capillary porosity; non-capillary porosity is related to granulation in most cases.

Henin (6) has used a penetrometer to obtain an index of tilth. This instrument measures the force of penetration of a metal point into the soil, as a function of depth. The force required for penetration is considered a measure of tilth. Blair (2) has proposed that compressibility be used to determine tilth. The more a soil can be compressed, the better the tilth; the amount of compression in this case denotes the extent of looseness.

The methods of Henin and Blair measure both consistency and structural effects. Penetration and compressibility are primarily dependent upon soil consistency but vary with structure. In other words, for a given soil structure, penetration and compressibility will vary with consistency. Therefore, although these types of determinations

offer considerable possibilities for obtaining an overall measure of tilth, they have not been characterized adequately enough as yet in terms of other soil properties.

Methods based upon granulation and porosity seem to offer the most promising means for characterizing soil tilth. Determinations of porosity must include not only the total porosity but also the relative distribution of the large and small pores. Cores of soils in their natural structure should always be used. The ideal technique would be to obtain a complete pF curve (size distribution of pores) on each core sample. In case the obtaining of a complete curve is not possible, sufficiently reliable results may be had by determining the non-capillary porosity from the difference in the air capacity at zero tension and at a pF of 1.5 to 2.0. A clod or aggregate analysis of the seedbed will give valuable supplementary data to help interpret the porosity results. Such an analysis may be made either by dry sieving, sieving in benzene or sieving in water. The stability of a given tilth condition can probably be estimated from the differences obtained between sieving in water and in benzene.

Further experimentation is necessary before thoroughly reliable means of characterizing tilth will be available. The ease of penetration and compressibility must be evaluated in terms of known structural conditions, as evidenced by the type and stability of the granules and the distribution of the pores relating to this granulation.

PLOWING

The plow has always been considered the most essential tillage implement in the production of crops. Even the most primitive agricultural civilizations used some type of implement, however crude, to perform part of the functions of the plow as it is now developed. It is true that in many instances the process of plowing included only the loosening of the soil so that seeds could be planted.

The development of the plow down through the centuries, in relation to increasing knowledge of soils and related sciences, is unusually interesting. The reader is referred to the excellent discussion of this subject by Keen (8b), who has traced the outstanding developments in the design of the plow and other tillage implements from the time of the Norman conquest to date, particularly under English conditions.

Mechanical Functions of the Plow. Nichols (14, 18) has listed the mechanical functions of the plow, in the order in which they occur during the passage of the plow through the soil, as consisting of the cutting loose, granulating and inverting of the furrow slice, and the

turning under of residues, weeds and green manures. Thus, from a purely mechanical point of view, the plow has been perfected to turn down surface material and to cause a loosening of the surface soil as a result of the breaking loose and inversion of the furrow slice. In the absence of surface material the primary function of the plow is to sever and granulate a given volume of soil, which is usually referred to as the furrow slice. The cutting loose of the furrow slice takes place at the edge and shin of the plow share. Granulation* occurs throughout the entire moldboard, although a considerable part of the granulating action may take place at the front portion of the plow surface. Lifting and inverting the furrow slice take place through the length of the moldboard. The completeness of coverage of materials depends upon the success with which the furrow slice is lifted and inverted.

Reasons for Plowing. Various reasons have been advanced for the necessity of plowing. In light of the type of implements used in primitive agriculture, perhaps the first reason for plowing was the loosening of the soil, so that seeds could be planted and crops would have a friable layer in which to grow. The next step was the introduction of a curved moldboard, for the primary purpose of turning under surface residues and live vegetation. The desire to plow under surface material as well as to loosen the soil has been reflected in the reasons usually advanced for justifying the need for the plowing operation. The object of turning under organic matter has been to get it out of the way for seedbed preparation and subsequent tillage operations and to increase its rate of decomposition. A satisfactory granulation of the furrow slice has been viewed as the first essential step in the producing of an ideal seedbed for plants.

On the basis of the recognition of this need for plowing, considerable progress has been made to develop plows that will permit the effective turning under of organic matter and that will accomplish a satisfactory granulation of the furrow slice. Consequently, there are plows for different soil and vegetation conditions. Sod plows have long sloping moldboards, so as to achieve successful inversion of the furrow slice. Plows for light soils without sod cover have steep moldboards to accomplish the inversion of the furrow slice. Other variations may be found between these two types for different soil conditions.

Recently, there has been considerable interest in finding out if the older concept of the necessity of frequent plowing for the aforemen-

* The term "granulation" used in this chapter is the common terminology of agricultural engineers to denote crumbling action.

tioned purposes is entirely correct. Is it always essential to plow under the residues that are on the surface? Is it always necessary to break loose and invert the furrow slice before crops can be successfully grown? These questions have been asked during the search for a solution of the problem of the deterioration of soil structure during the tilling and cropping of soils; the assumption has been made that the plowing operation has been damaging to soil structure.

A partial answer to the first question may be found in the investigations of Duley and Russell (4) on the use of crop residues in the Great Plains. They have found that organic matter left on the surface is much more effective in increasing the infiltration of water and decreasing evaporation than the same amount of organic matter plowed under. The presence of organic matter on the surface prevents the deterioration of granulation that usually results from the beating action of raindrops. This fact, along with other recent developments, indicates that it may not always be necessary to plow for the sole purpose of turning under certain residues. At the Missouri Agricultural Experiment Station, for example, successful crops of soybeans and small grains have been obtained on land that has not been plowed for six years. The stubbles of the preceding crop (lespedeza for the small grains and barley for the soybeans) have been left upon the surface and the soil stirred sufficiently to provide a suitable seedbed. Thus, there is some evidence to suggest that less plowing per rotation may be practiced in the future.

W. R. Williams of Russia (24) has offered an interesting explanation of why we plow, which may be found to have considerable merit if investigated properly. After stating that "practical observations extending over many centuries have shown that every soil, which in the spring possesses a perfect and durable cloddy structure, loses it during the growing period," he has attributed this breaking down of soil structure within the plow layer to the mechanical process of soil manipulation, the action of rainfall and the biological destruction of soil organic matter during plant growth. On the other hand, the structure of the subsurface layers is restored by the root systems of those plants that have a granulating action upon the soil. Consequently, he has visualized that the main objective for plowing is the turning under of this immediate surface layer, which has lost its structure, and bringing to the surface the subsurface layer, which has developed granulation through the action of the roots of vegetation. Shallow plowing or loosening of the soil without inversion is not considered of much value from the standpoint of structure.

In light of recognized agronomic practices which emphasize the

need of sod crops in the rotation, one might interpret the concept of Williams as suggesting that the plowing under of sod crops in the rotation is an essential tillage practice for promoting and maintaining good structure. Complete experimental verification of this hypothesis is lacking, although certain investigations now in progress at various institutions should throw some light on this problem.

There is reason to believe that the plowing operation itself has not been the cause of the universally recognized break down in soil structure. Seedbed-preparation activities probably have contributed more to this deterioration than plowing. If this is true, then more attention must be given to the achievement of better seedbeds through the plowing operation. This point will be discussed more thoroughly in a later paragraph.

Draft Requirements in Plowing. As previously mentioned, over half the power requirements in crop production are associated with plowing and listing operations. Before analyzing the causes of the high draft requirement during plowing, it is necessary to obtain a general picture of the forces involved in the movement of the plow through the soil. The outstanding work of M. L. Nichols at the Alabama Agricultural Experiment Station, on the dynamic properties of soils involved in tillage, will serve as a basis for most of the discussion to follow.

THE PHYSICS OF PLOW ACTION. In order to study the nature of plow action, Nichols (14) mounted a small nickel-plated plow in a long rectangular box in such a way that the point, shin and heel ran beside a heavy plate-glass window in one side of the box. The movement of the plow and the reactions of the soil to the plow were observed and measured by making use of a binocular microscope, which was mounted outside the window. Nichols observed that differences in soils as well as soil structure and moisture caused variations in the type of reaction obtained.

The most significant result of this experimentation was finding out that the action of the plow on the soil was exactly opposite to the generally accepted theory. According to the usual theory of plow action, which has been used rather extensively to explain the granulating effects of plowing, the granulation is produced by the fact that the lower soil travels farther than the surface soil as the furrow slice passes over the curved plow surface. Consequently, the different layers in the furrow slice travel at different speeds and granulation is produced. This action is illustrated in Figure 64a.

Nichols found that the action of the plow is opposite to this theoretical action, in that granulation occurs at right angles to the

tangents of the curve of the share and moldboard and not parallel to them. This actual soil movement is illustrated diagrammatically in

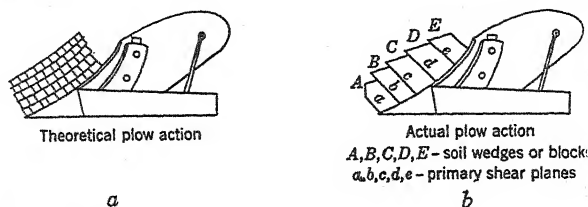


FIG. 64. Soil movements during plowing (Nichols, 14).

Figure 64b. The movements of the soil during plowing are explained as follows:

As the point of the plow advances in the soil, its bluntness (compared with soil particles) catches a part of the soil which it drives ahead in the form of a wedge. This soil wedge is compressed until its resistance to compression equals the resistance encountered in driving it into the soil. As this wedge advances through the soil there is a constant rolling motion of the soil along its sides due to the inter-locking of particles. This brings into account the force required to roll or slip soil over soil, which may be termed internal frictional resistance or "shear." "Shear" is usually preceded by a compression of the soil when the soil moisture is within what is ordinarily considered the plowing range. Obviously, a determination of the coefficient of internal resistance, compression values of the soil and the friction of soil on particular metal would make it possible to determine the value of sharp points on a plow.

As the point of the plow advances through the soil the inclined plane following it compresses the soil upward and forward. When the resistance to compression exceeds the shear value, the block of soil is sheared off and moves upward and forward as a solid unit. The force required for this action depends upon the pressure of the soil and the friction of the metal and the soil. The forward movement of this block of soil is, of course, due to the friction of the soil and metal and continues until the resistance of soil ahead of the block becomes equal to, or greater than, the friction between it and the metal when the soil is moved upward. When the plow shin is in the form of a curve, constantly increasing in steepness, the force required to move the soil up the plane is constantly increasing and the forward resistance of the soil must be greater to cause this more rapid movement. As the shin becomes more nearly perpendicular it will be noted that the surface applying force is nearly at right angles to the surface at the point. The pulverizing action of the plow is then due to the angles of the surface applying force being changed through the entire range of the curve of the plow from the nearly horizontal point to the vertical upper shin.

This may be more clearly explained by calling attention to the amount of forward travel in relation to the upward movement enforced on the soil by the plow's curvature. As the soil passes over various parts of this curve it is forced to slip over itself at a constantly varying rate. It should be noted that these force applications and resulting reactions are vastly different from, in fact being almost at right angles to, the forces of inversion to which is commonly attributed the pulverizing action of the plow. When the soil particles are cemented together by drying, this wave of shearing forces is still generally apparent, but due to areas of high and low bondage, the soil breaks into irregular lumps which grind over one another producing the same general result. However, in a cemented or dried soil the resistance to compression is high in proportion to the shear value and the pulverizing effect is limited to the fragmentation of the lumps. When a soil is in proper condition for tillage the stresses flow through the entire soil mass. On the other hand when a soil is so filled with moisture that there is free water present, or where the films are large enough to be easily broken, the waves of compression and shear produce the injurious puddling effect shown by the shining furrow slice.

On the basis of this explanation, it is seen that the granulating action of the plow is primarily due to the shear planes, along the shin of the plow, which extend across the furrow slice at an angle of 45° to the line of travel. The blocks of sheared soil are arranged in conformity with the shape of the moldboard. All shear planes move simultaneously across the moldboard, if the plow surface has the correct curvature. It is this simultaneous moving of the shear planes throughout the entire furrow slice that produces the desired type of granulation.

The importance of the correct curvature of the moldboard for good granulation is illustrated by the following reasoning, referring to Figure 64b:

As the plow advances, the soil slips over itself on shear plane *a*. To keep the soil slipping over itself in plane *b* at the same rate as the movement on *a*, the block *B* must travel forward at the same rate as the block *A*, plus an additional amount equal to the movement of *A*. To keep the soil block *C* slipping over *B* on the shear plane *c* at the same rate as the movements on *a* and *b*, the block *C* must move an amount equal to the combined movements of *A* and *B* plus the additional amount of the movement of *A*. The movement of the successive blocks of soil continues in this manner throughout the pulverization portion of the moldboard. This means that the movement throughout the soil mass on all shear planes is equal, that the total forward and upward motion is progressing with constantly increasing velocity as the soil slips up the plow and that pulverization by shearing is taking place throughout the mass of the

furrow slice. This is equivalent to the statement that the soil is being constantly and uniformly accelerated in a forward and upward direction as the plow advances.

Nichols and Reed (18) later analyzed the effect of the moldboard by studying a group of six different types of plows and their granulating influence upon sands, silt loams and clays. Even though these soils varied from a friable sandy loam to a tough, plastic clay, the

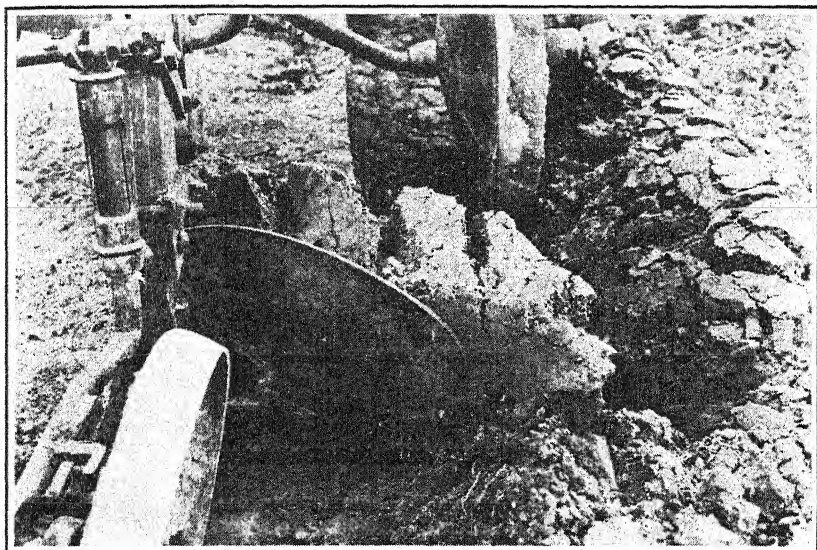


PLATE 4. Rear view of plow bottom, showing soil breakage along primary shear planes. (Courtesy of Bureau of Agricultural Chemistry and Engineering, U. S. Dept. Agr.)

general reactions to plowing were found to be of the same type. Working under field conditions, Nichols and Reed observed that the primary shear planes (45° angles forward from the plane of the shin) retained their same relative positions during their passage over the moldboard. At the point where the furrow slice stood in the furrow on its own furrow wall prior to its being inverted, the shear planes were parallel to the undisturbed furrow wall ahead of the plow, as shown in Plate 4. From this point the moldboard pushed the upper portion of the furrow slice forward and to the right to make it lie smoothly against the preceding one. It was rather common for these primary shear planes to retain their relative positions after the slice had been inverted, especially in the case of the heavier soils. The

significant fact is that these shear planes moved very little from the plane in which they were originally formed. Inversion of the furrow slice is accomplished by the rotation of the blocks of soil, formed by the shear planes, forward and to the right at an angle 45° to the line of travel until these blocks stand on end. They are then pushed over more forward and to the right, pivoting on the upper edge of the original furrow wall.

Nichols and Reed observed further that secondary shear planes were set up at an angle of 90° to the primary planes, which did not open until the furrow slice fell during the inversion process. These secondary planes are produced by the pressure of the advancing surface. The two types of shear planes are responsible for granulation of the furrow slice during the plowing operation.

DYNAMIC FACTORS INVOLVED IN PLOWING. After having determined the nature of plow action and established the types of forces involved, Nichols (14) investigated the interrelations of soil and plow design to the various forces involved in causing certain reactions to the passage of the plow. His classification of the different variables entering into the complex tillage problem is shown in Table 41.

TABLE 41
VARIABLES ENTERING INTO SOIL DYNAMIC STUDIES (NICHOLS, 14)
(Soil structure uniform; cementation zero.)

Primary Soil Factors (Measurable or controllable)	Design Variables (Controllable)	Dynamic Properties of Soil (Measurable)	Dynamic Resultants (Measurable)
Particle size	Kind of metal	Coefficient of internal resistance	Fragmentation
Colloidal content	Polish	(or shear value)	Arch action
Moisture (percentage)	Bearing area	Friction	Compaction
Apparent specific gravity (state of compaction)	Curvature of surface applying force	Resistance to compression	Shear
Organic matter		Cohesion	
Chemical composition of colloid		Adhesion	
		Moment of inertia	

This classification is the result of numerous ingenious laboratory studies with different soils at various moisture contents and with different structures. Variations in the kind and polish of metals as well as modifications in the design of the experimental "tillage tools" were used. The various measurable dynamic properties of the soil

will be discussed in the order in which they appear as the plow passes through the soil. Cohesion and adhesion, two exceedingly important factors that contribute to force relationships involved in tillage, have been discussed previously, in Chapter IV.

Resistance to Compression. The reaction of soils to applied pressures is one of the most important phases of the tillage problem. This is evident from the fact that most tillage implements apply pressure to the soil when they are used. Moreover, it has been shown that the first reaction of the soil to the forces applied by the plow is a resistance to compression, after which shear occurs. Nichols found that the percentage compaction increased with pressure. The curves were of a hyperbolic nature, in that compaction approached a maximum limit for a given soil at a constant moisture content. As shown in Figure 65, compression increases with moisture content to a maximum and then decreases as the amount of moisture increases. This is due to the increased ease of orientation of particles within the plastic range. In other words, plowing a soil in the plastic state will result in extreme compaction and puddling before shear reactions occur. The degree of compressibility increases logarithmically with decreasing plasticity number.

Shear. The shearing of soils is markedly different from the shearing of solids, with the possible exception of dried, puddled briquettes. The forces that produce shear are distributed by their action beyond a single plane. Under normal conditions of moisture, when one layer of soil is forced to slide over another, there tends to be increased contact and increased cohesion between particles rather than an immediate decrease in the holding together of the particles. There is a slight rolling effect as the coarser particles move over each other, which makes the area of shearing indefinite. Because of these factors, Nichols has defined shear as the "slipping of soil over soil"; it is characterized by "the internal resistance of a soil to any movement of its particles."

Internal friction and cohesion are the two types of forces that resist shear. If sufficient moisture is present, the slipping of one part of the soil over another may result in the immediate binding together of the particles by the cohesive forces in the water films; that is, after the shearing force is removed. The sheared area in this case may become more strongly cemented upon drying than other areas within the soil mass.

The results of Nichols (15) on the shear of non-plastic soils have indicated that the shear value increases in proportion to the colloidal material present. Sands had the same shear value at all moisture

contents. This should be expected if the colloid or clay content is responsible for variations in shear. The interlocking of the coarser particles of varying sizes, shapes and roughness is an important factor in the shearing of non-plastic soils.

Nichols (17) found that the shear value of plastic soils increased proportionally to the pressure applied to the soil at right angles to the direction of shear. The shear value increased with moisture to a maximum near the lower plastic limit (Figure 65) and then decreased to a very low amount at the upper plastic limit. In the light of the film theory of plasticity (see Chapter IV), this should be expected, since maximum film tension and cohesion occur near the moisture content of the lower plastic limit; the cohesion forces are very small at the moisture content of the upper plastic limit and flow is easily

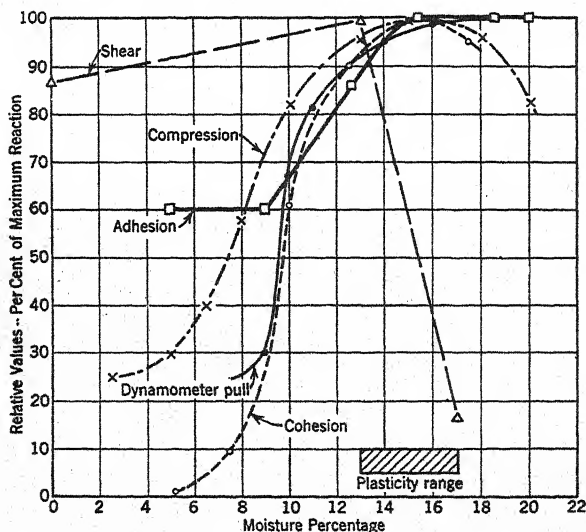


FIG. 65. The relation of dynamic factors involved in tillage to soil moisture, with special reference to the plasticity range. (The maximum value for each of these factors was taken as 100.)

produced. Once the soil becomes plastic, there is little internal friction resulting from the interlocking of particles. Consequently, shear is a function of the cohesion of moisture films beyond the moisture content of the lower plastic limit. As the moisture content approaches that of the upper plastic limit, shear gradually assumes the properties of viscosity.

The maximum value of shear was found to be proportional to the plasticity number, as should be expected if shear is a function of the

cohesion due to moisture films. It was also observed that the shear values of all plastic soils in the dry state were approximately the same under the same conditions of packing. This permitted the calculation of the shear value of plastic soils, up to the moisture content of the lower plastic limit, from data on the moisture content, the lower plastic limit and the plasticity number.

Soil and Metal Friction. Nichols (16) investigated the interrelationships of soil and metal by pulling a flat piece of the metal to be studied across a smooth soil surface. A calibrated spring balance was attached to the metal. The scale reading, divided by the weight of the metal pieces, gave the coefficient of kinetic friction, μ' . Different soils and metals were investigated at various moisture contents of the soil and speeds of pulling the metal across the surface.

The most important factor contributing to the friction between soil and metal was found to be the adhesion of soil moisture to the metal. When the soil began to adhere to the metal, through the medium of a water film, the observed coefficient of friction increased. The relation of the coefficient of friction to moisture content is shown in Figure 65. Nichols suggested certain fundamental relationships for characterizing the sliding friction that occurs between soil and metal. He divided the frictional resistance into four different phases, which were dependent upon the pressure of the metal and the structure and moisture content of the soil. They were defined as follows:

A. COMPRESSION PHASE. In a soil when the water does not adhere to the metal and when the bearing power of a soil is less than the pressure, that is, when the weight of the slider compresses the soil, the coefficient of sliding friction, μ' ,

1. Varies with the speed
2. Is proportional to the pressure per unit area
3. Varies with the smoothness of the surface and the materials of the surface.

B. FRICTION PHASE. When the bearing power of a soil is greater than the pressure per unit area and the soil water does not adhere to the metal,

1. The magnitude of the friction is proportional to the total pressure between the two surfaces
2. The value of μ' depends upon the roughness of the surfaces and the materials of the surfaces
3. It is independent of the area of contact
4. It is independent of the speed of sliding.

C. ADHESION PHASE. When there is enough moisture present to cause the soil to adhere to the sliding surface, but not enough to have moisture brought to the surface, then μ'

1. Varies with the speed
2. Varies with area of contact
3. Varies with the pressure per unit area
4. Varies with the surface tension of the film moisture, i.e.,
 - (a) It varies with the amount of colloidal material present
 - (b) It varies with the amount of water present
 - (c) It varies with the temperature and viscosity of soil solution.
5. Varies with the surface and kind of metal.

D. LUBRICATION PHASE. Where there is enough moisture present to give a lubricating effect, μ' varies

1. With the pressure per unit area
2. With the speed
3. With the amount of moisture and viscosity of the solution
4. With the nature of the surface and kind of material of which it is composed.

As previously shown, the structure and moisture content of soils determine, to a great extent, their resistance to compression. Consequently, these same two soil factors are responsible for the *A* and *B* phases in the friction curve. It was found that all dry soils, if compacted to the same extent, had the same coefficient of friction.

The moisture content is primarily responsible for the friction that is measured in the *C* and *D* phases. Moisture films between the soil and the metal are the determining factors, as shown in Figure 65; the maximum coefficient of friction is realized near the upper plastic limit. This should be expected since there are more films available for adhesion between soil and metal. For plastic soils, the maximum friction increases directly with the plasticity number. In other words, the greater the number of films, the larger is the adhesion between soil and metal. When the curve showing the relationship of the coefficient of friction to the plasticity number is extrapolated to zero plasticity, the friction value is exactly the same as that measured on the dry soils. This fact emphasizes the significance of adhesion-moisture film relations.

The friction curves of Nichols are quite similar to those reported by Haines (5). The former has gone much farther in the interpretation of soil-metal friction, with particular reference to implement design.

Relation of Dynamic Factors to Soil Consistency. In the previous discussion of soil consistency, attention was called to the fact that rather definite relationships had been established between consistency and the dynamic factors involved in tillage (1). These relationships are summarized in Figure 65. The values on the ordinate are only relative, since the purpose of this graph is to show the changes of the various dynamic factors with moisture content and to point out the relationship of these changes to the range of soil plasticity. Absolute values may be obtained by referring to reports of the original investigations (1, 16, 17).

As pointed out in preceding paragraphs, maximum cohesion takes place at the lower plastic limit. Film tension is greatest at this point. The range of maximum compressibility coincides with the plastic range. The soil can be compressed to its greatest extent at moisture contents that permit easy orientation of the particles. When the moisture content is higher than that of the plastic range, so much water is present that compressibility decreases. Maximum shear occurs at the lower plastic limit and decreases to practically zero at the upper limit of plasticity. This relationship is obvious in light of the film theory of plasticity. Maximum adhesion, as evidenced by values of the coefficient of friction, takes place near the upper plastic limits, for reasons previously mentioned.

After having established the fact that cohesion, adhesion, compression, shear and soil-metal friction affect the reaction of the soil to the plow, Nichols measured the overall effects of these factors on the force required to pull an implement through the soil. He compacted soils at varying moisture contents in a small glass box, slowly pulled chisels through the soil and measured the reactions obtained. A small dynamometer was attached to each chisel so that the actual pull in pounds could be recorded. The granulation of the soil ahead of the chisel was observed through the walls of the glass box. He found that the primary shear planes that were present with the small plow also appeared ahead of the chisels. The normal shear angle of soil reaction to the plow was about 45° . This same angle was obtained with the chisels regardless of the angle or curvature of the chisel; both 45° and 90° chisels were used.

An analysis of the effect of moisture on the resistance to passage of the chisels through the soil showed that maximum pull occurred within the plastic range, slightly above the lower plastic limit. Since the primary factors affecting the resistance of the soil to the passage of tillage tools are cohesion and adhesion, it is clear why the maximum resistance should be within the plastic range. Maximum pull increased

logarithmically with the plasticity number. From the slope of the curve below the lower plastic limit (Figure 65), one can readily see why it is essential to plow within the zone of friability if draft is to be decreased.

On the basis of these relationships between soil consistency and the dynamic factors involved in tillage, Nichols and his associates have arrived at a scientific basis of plow design for various soils. It is not the purpose of this book to discuss this phase of Nichols' investigations.

DYNAMOMETER STUDIES ON DRAFT. From an analysis of the forces involved in the breaking, granulating and inverting of the furrow slice, it is seen that the forces involved in compression, shear and soil-metal friction contribute to the draft of the plow. In addition, there is the work required to lift the furrow slice; this is but a small portion, however, of the total draft. Moreover, it has been shown that the draft, or dynamometer pull, for uncemented soils increases for any given condition with the moisture content to a maximum slightly above the lower plastic limit. In the case of compacted soils in the field, one should expect at least two maxima in the draft-moisture curve. If such soils are plowed when very dry, the cohesion between the dried particles will be high, as will be the dynamometer pull. As the soil becomes sufficiently moist to provide a water film thin enough around most of the particles to make the soil friable, the draft should decrease. The dynamometer pull will then rise to its second maximum when the soil is wet enough to be plastic.

As mentioned in Chapter IV, the plowing of dry soils of a hard consistency will result in the formation of numerous clods; plowing plastic soils causes puddling. Therefore, plowing a soil within the moisture range of the friable consistency not only achieves maximum granulation but also requires less output of power. This point is exceedingly important in all plowing operations.

The draft of plows under field conditions may be measured by means of specially designed dynamometers. The most notable example of such dynamometer studies has been the work of Keen and Haines (8) at the Rothamsted Experiment Station. They perfected a special dynamometer for soils work and then studied soil uniformity in relation to draft, as well as the effect of soil treatments on dynamometer pull. In their analysis of the variation in draft throughout a field, points of equal dynamometer pull were connected on a graph to form "isodynes." Their isodyne maps, therefore, contained "contour" lines of equal dynamometer pull just as a topographic map contains lines of equal elevation.

Perhaps the most significant feature of their findings at Rothamsted was the rather obvious heterogeneity of fields that should have been uniform, in light of their past treatment. The highest resistance was over 40 per cent higher than the lowest recorded. Interestingly enough, the location of the isodynes did not change with season. They were definite characteristics of the different portions of the field.

Only small differences were found between the various plot treatments. Annual applications of 14 tons of manure per acre showed some decrease in draft. The most significant decrease in soil resistance was accomplished by an exceptionally heavy application of chalk. The percentage reduction varied between 6 and 13 per cent. Applications of 5 tons per acre had little effect.

The roots of live vegetation play an important part in changing the dynamometer pull during plowing. Work is required to cut or tear loose the numerous roots within the plow layer. This fact was evidenced in dynamometer trials at the Ohio Agricultural Experiment Station, when the limed end of a series of plots required more draft to plow than the unlimed. This increased draft was not due to the direct effect of lime on the soil but to the fact that liming had made possible a luxurious growth of a legume-grass mixture.

By way of summary and repetition, since the plowing operation is the most costly tillage practice in crop production, more attention should be given to plowing at the right moisture content with the correct type of plow. Also, it may be possible to decrease the number of plowings per rotation.

PREPARING THE SEEDBED

The seedbed is usually considered as the plow layer which has been so prepared that planted seeds will readily germinate and the young plants will have a satisfactory physical condition for root development and growth. Unfortunately, the use of the term seedbed has emphasized the importance of preparing the few surface inches so that the seeds will germinate and emerge. Insufficient emphasis has been placed upon the structural qualities of the plowed layer that will permit and promote effective growth after emergence. In light of this weakness in terminology and concept, Slipher (21) has suggested that the term "rootbed" be used instead of seedbed. Rootbed would imply a satisfactory structure not only for germination and emergence but also for the complete development of the plant. Although this suggestion has many excellent advantages, the usage of rootbed is as yet rather limited. In light of this fact, the term seedbed will be employed

throughout this discussion, but it will be considered from the same point of view as rootbed.

REQUIREMENTS OF AN IDEAL SEEDBED. The structural condition of the seedbed is one of the keys to the successful growth of crops. The plant requires optimum air-water relationships from the time the seed is planted until the crop is harvested. Unfortunately, so much attention has been given in the past to the fertility aspects of crop production that the physical requirements of the plant have been seriously overlooked.

In many instances, seedbed preparation has been considered a necessary evil, prior to the planting of crops, and farmers and agriculturists alike often have not sought to achieve a satisfactory objective in preparing soils for crops. For example, seedbed preparation has almost always been considered a tillage operation that begins after the soil is plowed and ends after the seeds are planted. This is far too limited. Building a favorable seedbed for the growth of crops more properly begins prior to plowing and continues long after planting. In other words, the stable granular structural condition of the soil, which is a primary prerequisite for obtaining good tilth in the seedbed, must be produced before plowing. The plowing operation simply permits a falling apart of the furrow slice into granules that should be already present. A good job of plowing expedites the process of fitting the soil for the crop; a poor job increases the number of supplemental tillage operations necessary to complete a good seedbed.

Not only does preparation of the seedbed begin with or before the plowing operation but it also extends throughout the growing season. That is, a seedbed in good tilth at the time of planting the crop must be kept in this condition, as nearly as possible, during the growth of the crop. Consequently, the early operations in seedbed preparation will register enduring effects throughout the growing season. Cultivation of crops after emergence only tends to supplement the fitting operations, prior to planting, in maintaining good tilth for plant growth.

Slipher (21) has proposed certain basic requirements of the seedbed (rootbed according to his terminology) that emphasize the ideas just mentioned. They have been slightly modified as follows:

1. *Permit the rapid infiltration and satisfactory retention of usable rainfall.* With the exception of the sandy types, most soils have a sufficient water-holding capacity. However, the presence of a thin compacted layer at the immediate surface reduces the amount of infiltration, or intake of rainfall, and enhances surface runoff. This runoff, which may reach as high as 50 per cent of the

rainfall for certain torrential summer storms in the cornbelt, restricts the amount of water that may be available to the plant. It is significant to note that a compacted layer one-fourth of an inch thick at the surface of the soil may determine the rate of water movement throughout the entire profile.

2. *Afford an adequate air capacity and a ready exchange of soil air with the atmosphere.* Adequate aeration is essential for the seed, as well as the growing plant, to carry on normal respiration activities. Moreover, the aerobic microbiological processes depend upon a sufficient supply of oxygen. Thus, root development and normal growth are definitely related to aeration. Poorly aerated soils, however fertile from the standpoint of nutrients, are not productive. Fortunately, the same physical qualities of granulation that permit the rapid intake of rainfall also favor adequate aeration.

3. *Offer little resistance to root penetration.* Although factors other than compaction within the seedbed may be more responsible for the development of extensive root systems, nevertheless, any physical resistance to penetration tends to hinder the normal growth of roots. A good granular structure provides a more suitable medium for root development than a fine-grained compacted seedbed.

4. *Resist erosion.* Granular soils resist the dispersing action of raindrops. Consequently, the structure of the surface horizon is maintained and runoff is reduced. Smaller runoffs cut down soil losses due to washing.

5. *Facilitate the placement of surface residues.* The thorough distribution of organic materials throughout the soil mass accelerates their decomposition and conversion to stable humus compounds. This complete mixing tends towards better granulation than the mere covering up of residues on the bottom of the furrow.

6. *Provide stable traction.* A granular soil offers more stable bearing to tillage implements under a wider range of soil moisture than a fine-grained compacted soil. Moreover, cultivation for weeds can be accomplished more easily and effectively; the frequency of cultivation is reduced.

If these are the characteristics of good tilth, the question arises as to possible means of achieving the desired tilth in the seedbed. This question necessitates a concept of the seedbed profile, together, with the means of obtaining and maintaining such a profile through tillage operations.

THE SEEDBED PROFILE. In the old type of seedbed, the few surface

inches were reduced to a powdery condition as a result of overworking with tillage implements. Rollers, drags and other pulverizing types of tools were used to perfect this fine condition. Below this thin layer of finely divided soil, there was little disturbance of the plowed layer. The whole layer remained somewhat coarse and full of large air pockets. Removal of these air pockets was attempted by reaching them from the surface. Consequently, the structure of the immediate surface was destroyed in trying to establish better contacts between the surface soil and subsoil. This dusty surface is the antithesis of good tilth. The finely divided soil runs together easily under the impact of raindrops which rapidly reduces the rate at which water and air can enter the seedbed.

The ideal profile should be the old one inverted. That is, the lower part of the seedbed should contain the finest granules and possess the firmest degree of settling. The coarseness of the granules, in most cases, should increase as one approaches the surface. The immediate surface should consist of distinctly coarse granules to absorb the shock of the impact of raindrops and thereby preserve an open structure. It is true that the granules should not be so coarse in the vicinity of the seed that sufficient moisture cannot be had for the purpose of germination. On the other hand, the granules should not be so fine that inadequate aeration will prove to be the limiting factor. But what are the limits of coarseness or fineness? This question cannot be answered satisfactorily in light of existing data. Yoder (26) found that the emergence of cotton plants from seedbeds with different degrees of granulation was most rapid when about half the granules were $\frac{1}{8}$ to $\frac{1}{4}$ inch in diameter and the other half were smaller. A mixture of granules of varying sizes showed the best effects on stand.

This ideal type of seedbed aims at keeping the soil receptive to rainfall and favorable to air exchange. Its firmness underneath establishes sufficient contact with the subsoil to permit easy root development downward and an optimum movement of water within the soil profile. It must be realized, however, that various crops require different types of seedbeds. In some instances, the success or failure of the crop during germination and emergence may require a finer state of granulation than is satisfactory for the permanent tilth of the seedbed. Further experimentation is needed to establish the optimum seedbed requirements of various crops. Until such information is available it will not be possible to make specific recommendations for a given crop. However, it is certain that any general attempt to discourage the old type of seedbed will be reflected in better soil tilth for crop production.

TILLAGE TO IMPROVE TILTH. All tillage operations should aim at the production and maintenance of good tilth. Agronomic practices should also be designed to take advantage of the structure-forming qualities of a well-developed root system and organic matter. Therefore, plowing should be visualized as the first tillage operation in the preparation of the seedbed. Since there are many reasons to believe that the major destruction of soil granulation takes place by surface tillage operations after plowing, the importance of plowing in obtaining good tilth should be apparent. If the plowing operation produces a fairly satisfactory seedbed, that will require only a small amount of surface tillage, then good plowing will aid materially in the preservation of good tilth. If the plowing operation produces a poor type of seedbed, that requires a considerable amount of surface preparation, then poor plowing will result in the gradual deterioration of soil structure.

The success of the plowing operations in obtaining good granulation of the furrow slice depends primarily upon the granular nature of the soil and the moisture content at which the soil is plowed. There is an optimum range of soil moisture for each soil that permits the most effective results in plowing. This is the moisture range of the friable consistency. If plowed when it is not friable, a soil will usually be left in a cloddy condition. The breaking down of such clods requires many surface tillage operations, such as disking, rolling, harrowing, for a quick preparation of the seedbed. Fall or early spring plowing is often practiced on heavy soils that are difficult to find at the right moisture content. Nature then causes a breakdown of the clods by alternate freezing and thawing or wetting and drying.

Nichols and Reed (18) have observed that different physical conditions of the soil at the time of plowing materially affect the action of the plow upon soil granulation. They have described six different soil conditions and the type of reaction to the plow under each. They are as follows:

1. *Hard cemented soils.* This is a condition which is frequently experienced when soils are plowed too dry. The normal packing of the soil is magnified to a considerable extent by the shrinkage due to the rather severe drying. The soil has a harsh consistency and the cementation effects are large. Such soils break into large hard blocks; there is no definite arrangement of the planes of fracture. Plowing soils in this condition makes the preparation of the desired type of seedbed most difficult. Even after the slaking action of rains, considerable surface fitting is required. The seedbed is generally a mixture of clods and dust.

2. *Heavy sods.* Well-developed root systems of grasses form a rather dense mat in the surface of the soil, which tends to hold the furrow slice together. Even though it appears that the normal shear planes do not develop under these conditions, Nichols and Reed have found that careful examination will show the presence of these planes below the zone of greatest root concentration. It seems that rather good granulation is obtained, even though the furrow slice is bound together by the roots. Decomposition of the roots, even to a small extent, results in an ideal granular seedbed. If sods are plowed too wet, the amount of this granulation is decreased; sods usually are easy to prepare into a good seedbed if the plowing operation is performed far enough in advance of planting. This is more true of grass sods than of legumes. Stiff sods require gently sloping moldboards in order to obtain successful inversion of the furrow slice.

3. *Compacted surfaces.* A compacted surface or crust on soil, that is somewhat loose underneath, does not break regularly during plowing. Blocks of the surface layer are lifted up by the shear planes and shoved to one side. Granulation of the surface crust is difficult to achieve during the plowing operation; however, such a condition is rather uncommon.

4. *Freshly plowed soil.* The replowing of soils is not a common practice. Experience has shown that the soil must settle and offer a certain resistance to compression before the plow will function properly.

5. *Push soils.* Some soils in their natural settled condition react to the plow similarly to recently plowed soils. Adhesion to the moldboard is fairly high and cohesion within the furrow slice is low. Consequently, there is no slipping on the moldboard and the furrow slice is pushed to one side rather than inverted. Such soils tend to exhibit poor structural properties both before and after plowing.

6. *Normal condition.* This represents that friable state of soils in which maximum granulation is obtained and where the physical action of the plow is normal. With the possible exception of dense sods, this condition can be found in most agricultural soils, if they are plowed at the correct moisture content, and if previously protected by a cover of residue or live plant canopy.

Plowing, therefore, is a far-reaching tillage operation in the preparation of the seedbed. The first fundamental principle of a successful plowing operation is to plow at the right moisture content. Some

efforts have been made, in the form of attachments to the plow, to process the furrow slice as it is turned over. If such attachments can achieve better tilth in the seedbed without a surface manipulation of the soil, they will be of great value to the farmer. The second principle of good plowing is to plow those soils that are difficult to prepare, because of such inherent physical characteristics as poor structure and high clay content, far in advance of planting. Fall or early spring plowing will allow natural agencies (freezing and thawing, wetting and drying) to provide a granular structure in the surface.

The third requisite is to use the right type of plow to produce the granulation that is desired. Among the types of plows may be mentioned the rotary plow, which makes use of a fast-revolving drum, to which a number of tines are attached, to effect the so-called "pulverization." This plow is being used to a considerable extent in the horticultural industry for preparing seedbeds for vegetables. Although little experimental data exist concerning its effect upon the soil, the principle of its action is in direct opposition to the fundamentals of stable structure within the seedbed. Keen and his coworkers (9) found that rotary tillage caused earlier and better germination as well as improved early growth. The final yields, however, were either no better or distinctly inferior to those obtained on land prepared by the conventional moldboard plow.

Surface tillage operations in the preparation of the seedbed usually consist of disking, harrowing, leveling and compacting. The type and number of these operations vary with the supposed condition of the soil, the nature of the crop to be planted and the opinion of the farmer. It is not the purpose of this discussion to portray the relative merits of each of these implements. It is important to point out that, irrespective of the type of implement used, the larger the number of surface manipulations, the greater will be the eventual breakdown in granulation. It is true that the disk and harrow will loosen a compacted surface so that it is friable. This friability, however, must not be mistaken for the stable type of granulation that is formed through the biological agencies of root development and organic-matter production. In the majority of cases, tilth produced through surface manipulations is only temporary and superficial, however essential. Therefore, surface tillage operations should aim at producing the desired tilth with the least agitation. This will not only be best for endurance of tilth but also will decrease the cost of tillage operations.

In respect to those tillage tools that aim to crush or compact the soil, it should be pointed out that these operations are of questionable merit. It may be essential to compact a soil in order to establish a

firm seedbed for the germination of the seed. When that is essential, the immediate surface should not be left in a compressed state. If it is, a heavy rain will almost always cause a rather hard crust to form. The ideal seedbed should have rather coarse granules at the immediate surface, not finely powdered, compressed soil. If packing around the seed is essential to the good germination of a particular seed, several of the modern suggestions of subsurface compression at the time of planting may have certain possibilities. In many instances, it is possible that the benefits of compacting, as evidenced from increased germination, may be offset by later deleterious effects on the growth of the crop. Wollny (25), for example, found that rolling increased the speed of germination for a large number of farm crops but decreased the yield. Yoder also observed that the seedbeds which gave most rapid germination were not the ones that produced most cotton.

The packing of soil has always been explained as necessary for increasing temporarily the amount of water available to the seed or plant. When moist soil is squeezed in contact with a seed or roots, there is more soil per unit volume and, consequently, more water. The compaction of soil to increase capillary movement is very questionable in light of the slow movement of capillary water. If the soil were sufficiently moist so that compaction would increase capillary movement, the compression of such soils would exert a puddling action and tilth would be impaired. There is the possibility, however, that the packing of soils at low moisture contents may increase the availability of the rainfall from small showers during prolonged dry weather. This effect would be the same as making more water available per unit volume.

CULTIVATION

Cultivation usually refers to the tillage operation of manipulating the soil after the seed is planted or the young plant has emerged. There has been considerable disagreement in the past concerning the objectives of cultivation. Most of the disagreement has been the result of attempting to apply too widely the experimental facts obtained from a given situation.

Objectives of Cultivation. As a result of the classical experiments of King (10), on the effect of a dust mulch on the evaporation of water from soils, the generally accepted reasons for cultivation have been to conserve soil water. Emphasis has been placed (1) upon cultivation to control weeds so that they do not compete with the crop for water and nutrients and (2) upon the formation of a dry, dusty layer at the immediate surface to prevent evaporation losses. Numerous field

experiments since the original work of King, as well as a better knowledge of soil-moisture movement, have questioned the advantages of a dust mulch for moisture conservation. In light of these newer developments, the statement is often made that the purpose of cultivation is solely to control weeds; that exclusive position also is untenable.

When one analyzes thoroughly the pros and cons of cultivation, it seems apparent that the fundamental reason for cultivation is to maintain the seedbed in good tilth during the growth of the crop. This implies much more than simply controlling vapor losses from the soil, either by removing weeds or by the formation of a dust mulch.

One of the most significant discussions on reasons for tilling the soil was given by Johnson (7) of Connecticut in 1877. At this early date, he pointed out that tillage (1) gives the crop a well-aerated and healthy rootbed and (2) conserves rain against drought. The conservation of moisture is accomplished by increasing the infiltration capacity of the soil for water and by decreasing the losses through evaporation. He emphasized the significance of the compacting effect of raindrops on the soil.

In spite of the general impression that King emphasized the necessity of maintaining a dust mulch for controlling moisture losses, it is interesting to note in King's publications that he apparently pictured the cultivation operation as having much broader influences than reducing vapor losses. For example, he stated that tillage aimed to "destroy and prevent the growth of weeds," to "develop various degrees of openness of texture," to "modify the movements of soil moisture and of soil air" and in some cases to "change conditions as to make the soil either warmer or colder."

The modern trend on the objectives of cultivation, therefore, is not so distinctly different from the earlier views advanced by Johnson, King and Wollny. They can be more forcibly emphasized, because of a broader knowledge of soil-plant relationships. In the following discussion of the objectives of cultivation, the various reasons will be treated in order of their importance, at least according to the opinion of the author.

INCREASING THE INFILTRATION OF WATER. If the major fundamental objective of cultivation is to maintain good tilth during the growth of the crop, rapid infiltration of rainfall and adequate aeration are two important reasons for cultivating crops. It is difficult to judge which of these is the more important; there is insufficient evidence upon which conclusions can be based safely. Therefore, let us discuss the infiltration problem first.

Musgrave and Free (12) found that cultivation of the surface

greatly enhanced the water intake of soils. For example, the infiltration rates of water into a field soil (*a*) in its natural packing, (*b*) cultivated 4 inches deep and (*c*) cultivated 6 inches deep were 0.85 inch, 1.77 inches and 1.87 inches per 15 minutes, respectively. The first

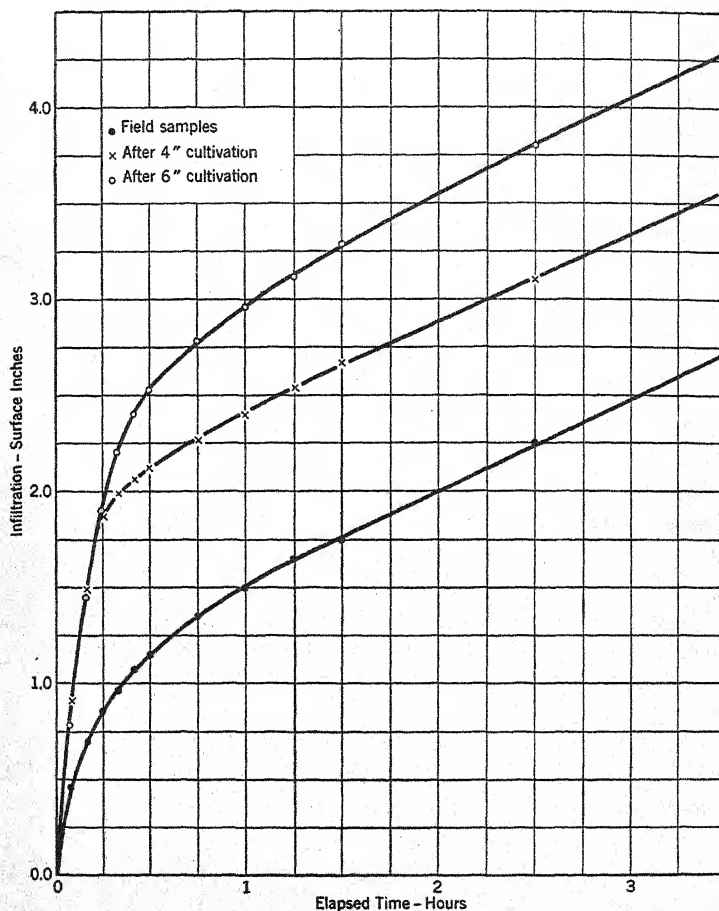


FIG. 66. The effect of cultivation on the infiltration capacity of Marshall silt loam (Musgrave and Free, 21).

hour rates for these same conditions were 1.49 inches, 2.31 inches and 2.93 inches, respectively. These results are illustrated in Figure 66. It should be noted that cultivation increases only the initial rates of infiltration. After one hour the curves are practically parallel. This means that loosening the surface soil and maintaining a high porosity at the surface cut down runoff by permitting the rapid intake of water

at the beginning of the storm. Once this cultivated layer is saturated, further infiltration is limited by the rate of percolation through the profile.

Neal (13) has demonstrated that the condition of the soil surface at the beginning of a rain is an important factor in determining the amount of runoff and erosion. His experimental results, which were obtained by artificial applications of rain on a small plot, are given in Table 42. The normal surface treatment consisted of cultivating to a depth of 4 inches and raking with an ordinary garden rake; the moisture content varied between one-fourth and one-half of the capillary capacity in the surface inch and between two-thirds and three-

TABLE 42
EFFECT OF THE CONDITION OF THE SOIL SURFACE ON RUNOFF AND SOIL LOSSES
(NEAL, 13)

Surface treatment	Cumulative soil and water losses at the end of the minutes indicated							
	10		20		30		40	
	Soil, lb.	Runoff, per cent	Soil, lb.	Runoff, per cent	Soil, lb.	Runoff, per cent	Soil, lb.	Runoff, per cent
Normal.....	0.20	3.6	1.09	30.3	2.42	45.3	7.61	65.3
Dry pulverized.....	0	0	0.12	9.1	0.48	20.9	2.84	42.6
Dry hard baked.....	1.86	41.2	5.50	63.2	8.71	72.7	16.68	83.6
Rough spaded.....	0	0	0	0	0.19	4.0	3.74	39.3

fourths of the capillary capacity throughout the rest of the cultivated layer. The very dry, pulverized surface was worked and dried throughout the upper 4 inches. The rough, spaded surface was prepared by spading to a depth of 4 to 5 inches after the soil was dry and allowing it to dry out without further treatment. The dry, hard, baked surface was prepared by allowing the soil to dry out after a heavy application of water. This type of surface simulated field conditions after an intense rain. Water was applied at the rate of 2 inches per hour on an 8 per cent slope.

It is seen that the hard surface lost most water as well as most soil. About 84 per cent of the total rainfall was lost. On the other hand, the rough, spaded surface, which simulated the condition of the soil in the field after cultivation, lost only about 39 per cent. Runoff on the hard, baked surface occurred 4 minutes after the rain started; that on the rough, spaded surface took place 26 minutes after the rain began. These experimental facts coincide with the data of Musgrave and Free, along with other results that have been reported.

It should be remembered that this was a very intense rain. Nevertheless, the cultivated plot did not lose much water or soil until after 30 minutes of rainfall. The hard surface permitted 10 times more runoff and soil loss in 10 minutes than the open soil did in 30 minutes. This fact emphasizes the importance of making the soil receptive for the rather intense rains that fall during the summer months. In many instances, these rains are of short duration. Unless the soil is in a physical condition to absorb them, a large percentage of the water will be lost to the plant.

Cultivation to increase infiltration is more important on rolling lands than on level terrain. Many of the conflicting results on the effect of cultivation on corn yields, for example, may be attributed to differences in the rate of infiltration of water between various soils.

ENHANCING AERATION. It was pointed out in Chapter VII that compaction of the soil led to decreased air permeability. Although little direct information exists concerning the extent to which cultivation increases the normal aeration of the root zone, there is reason to believe that the air supply to plant roots as well as to soil microorganisms is greatly augmented. Cultivation decreases the volume weight of the soil, which means that total porosity is increased. Moreover, the percentage of large pores (non-capillary porosity) is increased to a greater extent than the total pore space. Consequently, the total volume of air within the seedbed is raised.

Cultivation for better aeration is important on flat as well as rolling lands. As a matter of fact, it is usually more important on flat lands, since the standing of water on the soil tends to facilitate the formation of a compacted layer at the surface. Common experience has demonstrated that cultivation of compacted soils gives marked crop responses that do not seem to be associated with soil moisture. This is particularly true of such high air-requiring crops as sugar beets, potatoes and the like.

The loosening of compacted soils in the early spring speeds up bacterial decomposition and the mineralization of nitrogen. This increased bacterial activity is the result of a warmer and better-aerated soil.

CONTROLLING WEEDS. The seriousness of the weed factor in crop production is well recognized. Weeds compete with the crop for moisture, nutrients and light (in some specific cases). Consequently, every farmer attempts to control weeds and grasses by proper tillage operations. It should be kept in mind, however, that the tilth of the soil is often not concerned in cultivating for weed control. Weeds come up on soils in good tilth as well as on soils in poor tilth. Soils in good

tilth, however, can usually be cultivated more effectively for controlling weeds. Therefore, soils are cultivated with the major objective directed towards the weed rather than towards the soil. It is even questionable if weed destruction is technically a function of tillage.

The advent of the rotary hoe, weeder and other similar tools has made it possible to control weeds in the earlier stages of growth of crops without the usual type of cultivation. Such operations break the surface crust but have no other appreciable effect upon the tilth of the seedbed. Weeds and grasses should be killed as they emerge from the soil, if possible. At this stage, the tillage operation is fairly simple and effective.

REDUCING EVAPORATION. As previously stated, the work of King and others on the upward capillary movement of water has been interpreted as meaning that considerable losses of water through evaporation may take place unless the capillary columns in the soil surface are broken. This interpretation led to the theory of the dust mulch for moisture conservation. The formation of a dust mulch was considered one of the major functions of cultivation.

In 1917, Call and Sewell (3) in Kansas concluded, from their field experiments on cultivation in relation to soil moisture, that the formation of the dust mulch was only a secondary result of tillage for weed control. They studied the relative effects of removing weeds with the hoe, as compared with cultivation, on the moisture content of the soil. A summary of some of their results is given in Table 43. It was noted that the effect of cultivation varied with the type of season. It so happened that 1915 was a relatively moist year as compared with the other two years of the experiment. Consequently, there was little difference between the various treatments. During the dry years, however, the weed plot lost considerably more water than the others. There was no consistent difference between the scraped and cultivated plots. It is significant to note, however, that placing dikes around the plots in 1916 caused an increase in the moisture content of the weed-free plots. In light of these results, Call and Sewell concluded that cultivation, other than for the control of weeds, only conserved soil moisture by permitting a greater absorption of small showers.

Veihmeyer (23) studied the loss of water through evaporation under a wide range of conditions and found that a dust mulch did not produce a significant conservation of soil water. He observed that most of the water that is lost by evaporation leaves the soil before it is dry enough to cultivate so as to form a dust mulch. Similar results were obtained by Rotmistroff in Russia.

Recent developments in the knowledge of the capillary movement

TABLE 43

THE EFFECT OF CULTIVATION ON SOIL MOISTURE (CALL AND SEWELL, 3)

Date	Moisture content of variously treated plots			
	Weeds	Cultivated 3 in. deep	Cultivated 6 in. deep	Bare surface weeds scraped
April 13–Nov. 2, 1914:				
Av. H ₂ O—per cent.	16.16	19.52	18.08	18.98
Gain or loss—per cent. . .	−7.28	−3.31	−1.86	−0.89
April 15–Sept. 8, 1915:				
Av. H ₂ O—per cent.	22.50	23.40	22.90	22.40
Gain or loss—per cent. . .	−1.20	+0.70	−0.05	−0.70
May 29–Sept. 23, 1916:				
(a) Regular plots				
Av. H ₂ O—per cent.	15.94	19.09	16.49	18.14
Gain or loss—per cent. . .	−6.65	−2.39	−1.19	−2.50
(b) Diked plots				
Av. H ₂ O—per cent.	16.72	19.59	17.92	17.81
Gain or loss—per cent. . .	−3.65	+0.20	+2.37	+2.88

of water indicate that the flow of water in soils is very slow at moisture contents below the field capacity. As the soil dries out at the surface, the capillary conductivity across this dry layer is practically zero. Therefore, unless there is a water table close to the surface, which maintains a fair degree of saturation of the soil capillaries, the dust mulch should not be considered an effective barrier to evaporation.

Recognition of the fact that the old concept concerning the merits of the dust mulch is no longer tenable has led to an opinion that cultivation only for weed control is sufficient. This idea is also not tenable for the reasons stated in preceding paragraphs. It is dangerous to apply interpretations of the Kansas data to all situations. This fact is illustrated by some unpublished data from the Ohio Agricultural Experiment Station, as given in Table 44. Four different soils are included in this tabulation. Two, Paulding clay and Clermont silt loam, have notably poor structural properties. The former is a heavy non-granular clay; the latter is a silt loam with a single-grained structure. The Miami silty clay loam possesses a fairly good structure. The Brookston silty clay loam, on the other hand, has a rather stable

TABLE 44

EFFECT OF CULTIVATION ON THE YIELD OF CORN (OHIO DATA)

Soil type	Scraped	Cultivated 3 times
	bu./A.	bu./A.
Paulding clay, 1927.....	24.7	41.4
Brookston silty clay loam, 1927.....	68.2	74.1
Brookston silty clay loam, 1928.....	69.8	64.3
Brookston silty clay loam, 1929.....	94.0	90.4
Clermont silt loam, 1927.....	38.8	54.5
Clermont silt loam, 1928.....	34.8	36.5
*Brookston silty clay loam, 6-yr. av.....	72.3	69.8
*Miami silty clay loam, 6-yr. av.....	56.3	59.7

* Ohio Agr. Exp. Sta. Spec. Cir. 53, p. 37.

granular structure, and contains a considerable amount of organic matter.

The data in Table 44 show that the Paulding clay and Clermont silt loam are benefited considerably by cultivation. In other words, weed control was not sufficient for good yields. Both these soils are found on a flat topography. Runoff, therefore, does not seem to be the reason for the increased yields. It is highly probable that aeration effects have played a considerable role in the higher yields produced by cultivation. The slightly higher yields on the cultivated Miami silty clay loam may have been due to decreased runoff, since this soil type is found on gently rolling areas. Removing weeds apparently was sufficient for obtaining good yields on the Brookston soil. Cultivation had no effect upon corn yields; only one year out of three did cultivation cause higher yields. This should be expected from a knowledge of the physical properties of the Brookston. It is usually friable throughout the growing season. Moisture absorption and aeration are generally favorable. Under such conditions, about the only plausible reason for tillage would be to control the weeds. These data clearly indicate that the different objectives of soil cultivation must be changed according to season, soil and crop. The objective of maintaining good soil tilth is of major importance.

Frequency and Depth of Cultivation. The frequency of cultivation depends upon the necessity of maintaining the desired type of tilth and of controlling weeds. Unfortunately, those soil types that

do not have a stable structure are the ones that require most frequent cultivations. Those soils that have stable granules and could withstand the manipulative effects of frequent cultivations do not require as much tillage to maintain tilth. Therefore, it is apparent that the poorer the structure the greater will be the need for frequent loosening of the surface, and the more rapid will be the eventual deterioration of structure. These facts emphasize the necessity of providing good tilth in the seedbed by planning the rotation prior to the plowing operation. Seedbeds following sod crops have fairly stable granulation, at least during the first growing season after plowing.

The aforementioned Ohio experiments also have shown that six to seven cultivations were seldom as good as three cultivations. Other experimental data indicate similar relations (19). Russell and Keen (20) found that the number of cultivations, beyond those necessary for adequate weed and crust control, did not produce valuable results.

The depth of cultivation depends primarily upon the depth of compaction. On those soil types that normally exhibit good tilth, rather shallow cultivation may suffice. Deep cultivation seldom is profitable and may even prove harmful to the root system of the crop. Results of eight years of experimentation on the relation of the depth of cultivation to the yield of corn on Wooster silt loam (19) show the following: corn cultivated 1 to 2 inches deep, 64 bushels; cultivated 2 to 3 inches deep, 72.2 bushels; and cultivated 4 to 5 inches deep, 67 bushels. These data indicate that deep cultivation is less desirable than a relatively shallow stirring of the surface soil. It is probably safe to suggest that cultivating to a medium depth will be most satisfactory over a wide variety of conditions. Shallow or deep cultivation may be advantageous for certain specific cases.

In summing up the needs for cultivation, it should be emphasized that this operation should be performed at the correct moisture content. Since the basic objective is to produce good tilth, cultivating at moisture contents corresponding to the friable consistency will result in maximum granulation. One should also keep in mind that the granulating influences of cultivation are not permanent but tend to disappear under the influence of raindrops and surface manipulation. The degree of permanency of the granulation in the seedbed usually determines the frequency of tillage. Thus, a knowledge of the physical requirements of the seedbed for a given crop is essential in arranging a program of tillage.

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CHAPTER X

PHYSICAL PROPERTIES OF SOILS IN RELATION TO RUNOFF AND EROSION

THE ROLE OF THE SOIL IN THE HYDROLOGIC CYCLE

The hydrologic cycle refers to the disposition of rainfall from the time it is precipitated until it re-enters the air masses by re-evaporation to serve again as a source of precipitation. The hydrologic cycle is illustrated in Figure 67. It is noted that part of the rainfall evaporates

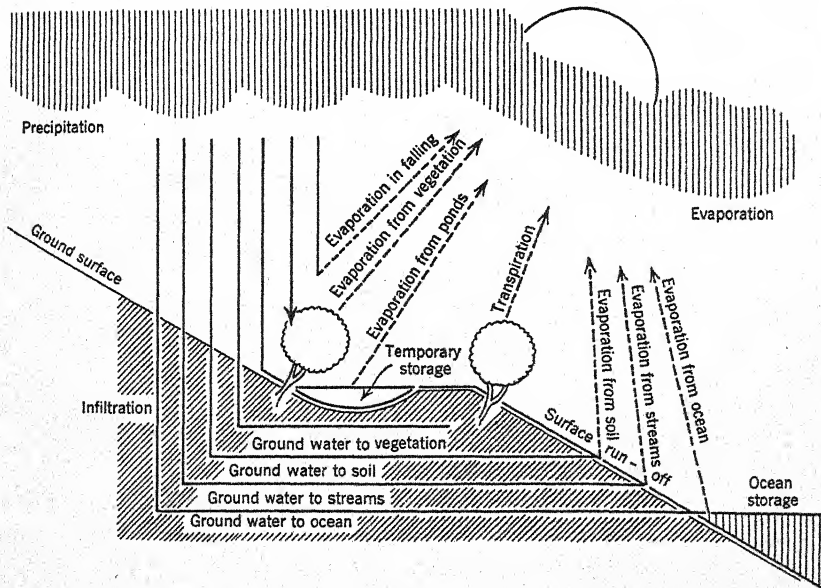


Fig. 67. The hydrologic cycle.

before it reaches the land surface, that is, it either evaporates while falling or after it is intercepted by vegetation. That part of the rainfall which reaches the surface of the soil either runs off or enters the soil or both. Runoff water may be impounded in lakes and reservoirs, or it may reach the streams and finally the ocean. Evaporation into the

atmosphere takes place from these lakes, streams and the ocean. During the process of runoff, soil erosion generally occurs.

Water that enters the soil may be returned to the atmosphere by direct evaporation, transpiration by vegetation or by evaporation from streams and the oceans after reaching these bodies of water through ground-water flow.

This brief resumé of the different phases of the hydrologic cycle points out that the soil plays an important role in the disposition of precipitated water. It is a complex variable in the whole problem of hydrology.

AN ANALYSIS OF THE PROBLEM OF RUNOFF AND EROSION

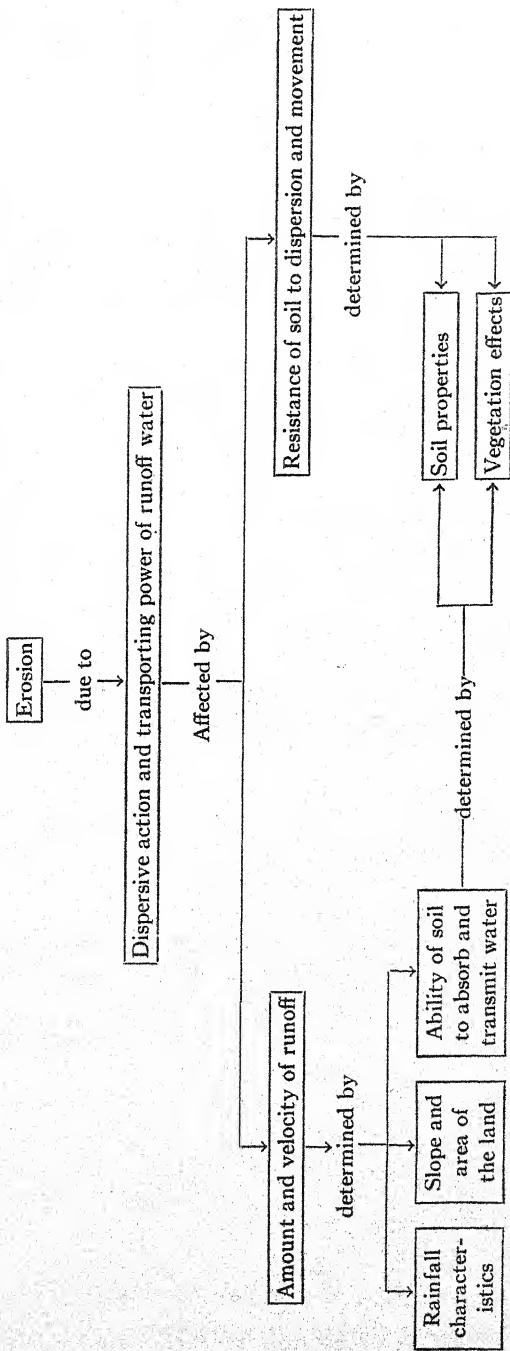
Water erosion is due to the dispersive action and transporting power of runoff water. If there were no runoff, there would be no erosion. On the other hand, if runoff water could be prevented from bringing soil into suspension as water travels across the surface, there would be no erosion. Unfortunately, there are certain rains which the most permeable soil cannot absorb. Consequently, there is runoff and usually erosion.

The dispersive action and transporting power of water are determined by (1) the amount and velocity of runoff and (2) the resistance of the soil to dispersion and movement. If the amount of runoff is small and its velocity is low, erosion will not be severe. If the total amount is large and the velocity is low, soil movement is not likely to be excessive. When both factors are high, erosion is generally serious. For any given characteristics of runoff, various soils will erode differently dependent upon the resistance that is offered to dispersion and movement. If the soil is protected by a good vegetative cover, only slight erosion will occur under excessive runoff. If the soil is not protected, there is usually some erosion for the smallest runoffs. The degree of soil movement will be dependent upon the ease with which the soil can be dispersed.

The amount and velocity of runoff are dependent upon (1) rainfall characteristics, (2) the slope and area of the land and (3) the ability of the soil to absorb and transmit water through the profile.

The relationship of water erosion to the various contributing factors is shown schematically in Table 45. It is seen that the erosion problem is made up of the following variable factors: (1) climate, (2) topography, (3) vegetation and (4) soils. Since man has played a considerable role in accelerating the erosion process and will portray an equally important part in controlling erosion, a fifth variable, the human factor, should undoubtedly be introduced.

TABLE 45
FACTORS AFFECTING SOIL EROSION DUE TO WATER



In a general way, therefore, the factors affecting soil erosion may be summarized according to the descriptive equation:

$$E = f(C, T, V, S, H) \quad (51)$$

in which the independent variables refer to the aforementioned factors. This equation can be descriptive only because of a lack of knowledge of the significance of the different variables, as well as the difficulty of expressing the variables in similar terms. Nevertheless, erosion is an explicit function of the independent variables cited; moreover, the different factors are implicitly related.

This descriptive equation is of interest from several points of view. In the first place, it is clear that a complete solution of the erosion problem depends upon an evaluation of each of the variables with respect to each other. Knowledge concerning the character of erosion as well as effective control methods can be advanced only by investigating the interrelationships of the factors mentioned. For example, on a given farm with a uniform soil and the same type of vegetation, the last three variables will become practically constants and erosion will vary as a function of the type of rainfall and the slope and area of the land. On the other hand, if the same rainfall characteristics obtain on different soils with the same slope and area of watersheds, erosion will be a function of vegetation and soil characteristics.

Second, this equation contains two types of variables, those that can be controlled and those that are directly uncontrollable. For example, climate, the degree of slope of the land and certain physical characteristics of the soil cannot be directly controlled. Their effects, however, may be modified indirectly, such as by the use of terraces to cut down the length of slope.

It is not the function of this book to discuss these erosion variables in detail, excepting the soil factor. Since there are sufficient climatic, topographic and vegetation effects that directly or indirectly influence the soil variable, it will be advantageous to call attention to several of the more important phases of these three factors.

The Climatic Factor. The major climatic factors that influence runoff and erosion are precipitation and temperature. Precipitation, of course, is the more important. Temperature affects runoff insofar as it contributes to changes in soil moisture between rains, determines whether the precipitation will be in the form of rain or snow and changes the absorptive properties of the soil for water by causing the soil to freeze to an appreciable depth.

Confining our attention to precipitation as rain, we find that the

amount, intensity and distribution of the rainfall help to determine the amount and velocity of runoff and the losses due to erosion. A large total rainfall may not cause excessive erosion if the intensity is low. Likewise, an intensive rain of extremely short duration may not cause much erosion because there is not enough rainfall to produce runoff. On the other hand, when both the capacity and the intensity factors are high in a given storm, both runoff and erosion will be serious.

These relationships are well illustrated by the results of Hays and Palmer (8), as given in Table 46. These data show that 2.6 inches of rain that fell with a very low intensity did not cause appreciable erosion. Also 0.9 inch of a very intense rain did not produce serious soil losses. However, the rain of July 5, which combined amount and intensity,

TABLE 46
THE RAINFALL FACTOR IN SOIL EROSION *

Date of rainfall	Amount of rainfall, in.	Maximum intensity of rainfall, in./hr./10 min.	Duration of rain	Soil losses, tons/acre
June 18 and 19, 1935.....	2.6	0.3	30 hr., 35 min.	0.4
July 5, 1935.....	1.9	2.8	1 hr., 52 min.	51.2
July 27, 1935.....	0.9	3.5	15 min.	2.2

* Land in corn, planted on contour on a 15 per cent slope. O. E. Hays and V. J. Palmer: Soil and Water Conservation Investigations. Soil Conservation Service. Progress Report, 1932-1935, Upper Mississippi Valley. Soil Conservation Experiment Station, LaCrosse, Wisconsin.

removed over 51 tons of soil per acre. Numerous other data could be given which show similar relationships of the capacity and intensity factors of rainfall to erosion. Suffice to mention, however, the work of Neal (21) which showed that erosion increased according to a power function of the intensity. For the soil and slope conditions of his experiments, erosion varied as the 1.2 power of the rainfall intensity.

The distribution of rainfall is exceedingly important in runoff and erosion problems. Distribution in respect to the amount of water in the soil as well as to the type and amount of vegetative cover is especially significant. An intense rain that falls on wet soil will have an entirely different runoff and erosion pattern from the same type of rain that falls on dry soil. Likewise, a storm that occurs during the seeding of a wheat crop, for example, will cause much more ero-

sion than a similar storm that occurs after the wheat has produced several inches of top growth.

The significance of rainfall distribution to runoff and erosion, with particular reference to the classical soil-erosion plots at the Missouri Agricultural Experiment Station (17), has recently been pointed out (3). Runoff and soil losses from these plots always presented perplexing results when analyzed solely from the point of view of the amount and intensity of the rainfall. For example, there was always more runoff and erosion in April than in May, although May had more rainfall and more intense storms than April. Moreover, more runoff and erosion occurred in September than in August, although the rains falling in August had a greater intensity.

Since an analysis of the intensity and amount of precipitation did not permit a complete picture of the significance of rainfall characteristics in these experiments, an attempt was made to evaluate the importance of the effective moisture content of the soil during precipitation. Inasmuch as no continuous record of soil moisture was kept, it was impossible to calculate directly the effective moisture in the profile. In light of this fact, it was necessary to estimate indirectly the relative effective moisture content of the soil for the various months by using precipitation and temperature data and calculating the precipitation-evaporation ratio. A modification of the method of Thornthwaite was used because he had shown the value of using monthly data for interpreting the rainfall characteristics of a region (24). The effective soil moisture was considered to vary approximately according to the ratio, $10^3 P/T^2$, where P is the mean monthly precipitation in inches and T is the mean monthly temperature °F.

When the factor of the moisture content of the soil at the time of precipitation was evaluated, it became evident why more soil and water were lost in April than in May. (See Figure 68.) Most of the April rains fell on wet soil. A similar explanation was possible for the August and September differences. The extremely dry soil in August was able to absorb a considerable portion of the torrential rains falling at that time. In September, the soil was considerably more moist.

From a practical as well as a technical point of view, it is essential to study the rainfall characteristics of a region in relation to the cropping systems. It is necessary to know whether the vulnerable points in a given cropping system coincide with the "intensity peaks" of the rainfall curve. If this happens to occur, there is always the desirable possibility of changing the rotation to one which is less vulnerable. For example, the farmers along the Missouri River in central Missouri

have always practiced a three-year rotation of corn, wheat and clover. Ordinarily, this should be a good soil-protecting rotation. It so happens, however, that the farms in this area have suffered severe erosion in spite of fairly good management. Erosion has been due to the rather strongly rolling topography and to the distribution of the torrential rains. The most severe peak in the rainfall curve occurs in August and September, when the farmers are preparing their land for wheat. A peak in the effective moisture curve comes in September. Consequently, sheet erosion has been severe.

The rainfall-distribution pattern for Columbia, Missouri, in relation

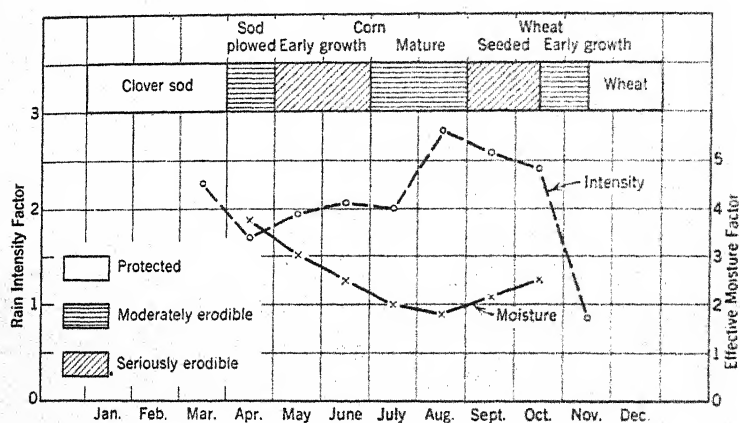


FIG. 68. The importance of rainfall distribution, with respect to the protection offered by crops, to erosion. Data from Columbia, Missouri. Rotation is corn, wheat and clover.

to the vulnerable points of a corn, wheat and clover rotation is shown in Figure 68. The most vulnerable year of this rotation is the one in which both corn and wheat are planted. Fair protection is provided by the wheat crop for the normal types of rain. Good protection is given by the clover sod. This rather simple illustration indicates the desirability of considering the vegetative protection provided by the crop in relation to the distribution of erosive type of rains.

The Topographic Factor. Slope characteristics are also important factors in determining the amount of runoff and erosion. Erosion is usually not a problem on extremely flat lands. As soon as the topography becomes slightly rolling, erosion begins to be serious. The degree and length of the slope are the two essential features of topography that are concerned in runoff and erosion. The uniformity

of slope is often important in determining the relative ease or difficulty of establishing suitable erosion-control practices.

Of the two characteristics of slope, degree and length, the former is usually the more important from the standpoint of the severity of erosion. This is indicated by the data in Table 47. It is seen that on slopes below about 10 per cent the amount of erosion approximately doubles as the degree of slope increases twofold. Diseker and Yoder (6) found that it more than doubled. Neal (21) observed that it depended somewhat on the intensity of the rain. These investigators found that the curve relating erosion as a function of slope for any given rain was slightly S-shaped. The losses from the steeper slopes did not increase in the same proportion as the losses from the more gentle slopes. Neal's data showed that erosion varied as the 0.7 power of the percentage slope. Most of the experimental data indicate that the degree of slope has little effect upon the percentage of runoff.

The effect of length of slope on erosion seems to vary considerably with the type of soil. This fact is indicated by the data in Table 48. These results show that for some soils, Muskingum silt loam and Vernon fine sandy loam, the length of slope has had no apparent effect on total runoff, even though the erosion increased with length. The longer slopes on the Clinton silt loam and Houston black clay had less runoff than the shorter ones. Apparently, these differences are associated with variations in permeability between the two groups of soils. Erosion on the Clinton soil increased with length of slope; on the Houston soil it decreased with length. The different behavior of these two soils in regard to erosion is probably due to the wide difference in degree of slope.

Musgrave (19) has offered some valuable suggestions as to the effect of length of slope on runoff and erosion. He has found that the intensity of rainfall greatly influences these losses. For example, as shown at the bottom of Table 48, runoff and erosion increase with length of slope for the high-intensity rains. The reverse is true for rains of low intensity. The Marshall silt loam is a highly permeable soil. With rains of an intensity only slightly higher than the infiltration rate, there is greater total infiltration with the longer slopes. With intense rains; there is less time for infiltration which results in more total runoff and a greater velocity of runoff. Moreover, the intense rains may cause a partial destruction of the structure in the surface of the soil which leads to more rapid runoff and greater erosion.

The data in Tables 47 and 48 point out the variations that may be expected in erosion on various types of slopes. They also emphasize

TABLE 47
THE RELATION OF DEGREE OF SLOPE TO EROSION

Soil type	Amount of rainfall, in.	Duration, hr.	Soil losses in tons per acre for the following slopes in per cent									
			0	2	4	5	8	10	12	15	16	20
Cecil clay *	20	164	0.4	13.3	40.1	67.1	91.3
	13.39	155	0.3	14.5	36.9	46.1	59.1
Putnam silt loam †	4	1	...	9.3	11.9	19.7	34.8
	3	1	2.1	3.1	6.9	11.5	13.7
	3	2	...	1.0	2.5	5.4	6.7	9.1
	9	3	...	9.6	17.8	26.4	33.7
Muskingum silt loam ‡	38.38	1934-1937 average	62.3	76.9	83.2

* Plot length, 50 feet; area, $\frac{1}{8}$ acre. See Diskier and Yoder (6).

† Plot length, 12 feet; area, $\frac{1}{100}$ acre. See Neal (21).

‡ Plot length, 76.9 feet; area, $\frac{1}{100}$ acre. See Borst and Woodburn (5).

TABLE 48
THE EFFECT OF LENGTH OF SLOPE ON RUNOFF AND EROSION

Location	Soil type	Degree of slope, per cent	Length of slope, feet	Runoff, per cent of annual rainfall	Erosion, tons per acre per year
Zanesville, Ohio * (4-yr. average), 1934-1937	Muskingum silt loam	12	36.3	39.5	72.4
			72.6	42.3	76.9
			145.2	39.2	91.6
LaCrosse, Wis. † (3-yr. average), 1933-1935	Clinton silt loam	16	36.3	20.8	62.6
			72.6	20.4	88.7
			145.2	18.0	100.9
Temple, Tex. ‡ (6-yr. average), 1931-1936	Houston black clay	4	36.3	16.5	27.1
			72.6	14.2	23.8
			145.2	13.6	21.7
Guthrie, Okla. § (5-yr. average), 1930-1935)	Vernon fine sandy loam	7.7	36.3	15.9	21.3
			72.6	15.3	25.5
			145.2	15.7	39.2
Clarinda, Iowa (3-yr. average), 1932-1934)	Marshall silt loam H = High-intensity rains L = Low-intensity rains	8	157.5	H 10.8 L 28.0	H 8.6 L 7.7
			315.0	18.0 16.5	18.3 6.3
			630.0	20.3 13.5	32.9 5.7

* H. L. Borst, Russell Woodburn. *Progress Report Northwest Appalachian Soil and Water Conservation Experiment Station, 1933-1937.*
† O. E. Hays and V. J. Palmer. *Progress Report Soil and Water Conservation Investigations Upper Mississippi Valley Soil Conservation Experiment Station, 1932-1935.*

‡ E. B. Deeter and P. L. Hopkins. *Progress Report Blackland Soil and Water Conservation Experiment Station, 1931-1936.*

§ H. G. Lewis and H. S. Riesbol. *Summary of Results Red Plains Soil Erosion Experiment Station, 1930-1935.*

|| G. W. Musgrave. *Trans. 16th Ann. Meeting Amer. Geophysical Union, 1935, Part II, pp. 472-478.*

that soil properties may play an important role in determining the type of runoff and erosion with varying slope conditions.

The Vegetation Factor. A good vegetative cover, such as a thick sod or a dense forest, offsets the effects of climate, topography and soil on erosion. This fact is particularly emphasized by the experimental results of the Federal Soil and Water Conservation Experiment Stations. A good grass sod has permitted less than one ton of soil loss per year on soils ranging from the highly permeable Marshall silt loam to the fairly impermeable Shelby loam, and on slopes varying from 4 per cent at Temple, Texas, to 30 per cent at LaCrosse, Wisconsin. Naturally, the agricultural production of various crops cannot be maintained by having all the land covered with trees and grass. But, even under conditions of cropping, vegetation effects play a significant role in controlling erosion.

The major effects of vegetation may be classified into at least five distinct categories. They are (1) the interception of rainfall by the vegetative canopy, (2) the decreasing of the velocity of runoff and the cutting action of water, (3) the root effects in increasing granulation and porosity, (4) biological activities associated with vegetative growth and their influence on soil porosity and (5) the transpiration of water leading to the subsequent drying out of the soil. It seems obvious, therefore, that a discussion of the soil factor in the erosion problem cannot be separated from the vegetation factor; they are interdependent.

INTERCEPTION OF RAINFALL. The interception of raindrops by the canopy of vegetation affects soil erosion in two ways. In the first place, part of the intercepted water never reaches the soil but is evaporated directly from the leaves and stems. This water, therefore, cannot contribute to runoff and is not a factor in erosion. Second, the vegetative canopy absorbs the impact of the raindrops and thereby minimizes the destructive effects of the beating action of the rain on soil structure. This particular effect had been emphasized by various early investigators, notably Schumacher, Wollny and Johnson. Wollny showed that the denser the plant cover, the less was the change in porosity of the soil during the growing season. Some of his data was given in Table 22, Chapter V. He contended that the dispersion of the soil by the raindrops resulted in a clogging of the pores and a diminishing of soil porosity. These results were later confirmed by Lowdermilk (14) and Hendrickson (9).

In 1880, Wollny (4) investigated the effect of plant cover upon the interception of rainfall and found that only 45 to 88 per cent of the total rainfall reached the land surface directly, depending

upon the type of crop and the number of plants per unit area. The results for ordinary farm crops are given in Table 49. It is interesting to note that corn, when planted thick, was the most effective inter-tilled crop for intercepting rainfall. Red clover and lupines were the most effective forage crops. These results are of the same order of magnitude as those reported by Haynes (7).

TABLE 49

THE EFFECT OF PLANT COVER ON THE INTERCEPTION OF RAINFALL

Crop	Per cent of total rainfall penetrating vegetative canopy					Haynes (7)
	Wollny (4)					
	No. of plants per 4 sq. meters					
	0	36	64	100	144	
Corn.....	100	62.9	60.7	57.0	44.5	68
Soybeans.....	100	88.4	78.2	65.9	64.3	65
Oats.....	100	78.5	78.4	78.9	93
Vetch.....	100	78.1
Lupines.....	100	57.9
Peas.....	100	87.8
Red clover.....	100	61.3
Alfalfa.....	57

These data help to explain the fact that an intertilled crop, such as corn, even when it is cultivated up and down the slope, reduces erosion considerably over that which occurs on fallow land. The total runoff between fallow and continuous corn is usually not very different. Undoubtedly, the canopy effect of corn is important in breaking up the fall of the raindrops. It seems safe to predict that future experimentation, on the degree to which the impact of raindrops is lessened by vegetation, will show that this phase of the interception of rainfall is more important than is generally recognized in the prevention of soil erosion. The investigations of Kramer and Weaver (13), in which they have found that the presence of the tops of vegetation decreases the rate at which a block of soil erodes under artificial rain, tend to substantiate this prediction.

DECREASING RUNOFF VELOCITIES. Any vegetative cover is an impediment to runoff water. A well-distributed, close-growing surface

growth not only will slow up the rate at which water travels down the slope but will also tend to prevent a rapid concentration of water. Reducing the velocity of runoff and preventing the concentration of this water lessen the cutting action of the water.

The fact that vegetation exerts a greater influence upon the velocity than upon the total amount of runoff is seldom appreciated. These effects are clearly shown in the data of Norton and Smith (22) in Table 50. These results point out that the difference between grass and corn with regard to rates of runoff is about twice as great as the difference between these two crops with respect to total runoff. Moreover, even though the total runoff from oats and bluegrass is ap-

TABLE 50

THE EFFECT OF VEGETATION ON THE AMOUNT AND VELOCITY OF RUNOFF
(NORTON AND SMITH, 22)

Year	Number of rains	Crop	Maximum rate of runoff, inches of rain per hour	Total runoff, inches of rain
1933-1936	50	Corn.....	0.67	0.57
		Clover and timothy....	0.27	0.38
		<u>Corn</u>	2.50	1.50
		<u>Clover and timothy</u>		
1933-1934	36	Corn.....	0.79	0.34
		Bluegrass.....	0.20	0.13
		<u>Corn</u>	3.95	2.60
		<u>Bluegrass</u>		
1935	19	Oats.....	1.23	0.52
		Bluegrass.....	0.53	0.50
		<u>Oats</u>	2.30	1.00
		<u>Bluegrass</u>		

proximately the same, the rate of runoff from the former was over twice as great as that from the bluegrass. When the rate of runoff is decreased by vegetation, there is more time for infiltration which reduces total runoff.

A good grass sod resists the cutting action of water. This is evidenced by the fact that sod-stabilized terrace outlets are among the best and that many terraces are emptied into old pastures. The

use of grass waterways to conduct water down the slope testifies also to the resisting qualities of sod to runoff. Therefore, it should be clear how a vegetative cover helps to determine the soil's resistance to dispersion and transportation.

ROOT EFFECTS. The root effects of vegetation in the production of a stable granulation and greater porosity have been thoroughly discussed in Chapter V. The binding influences of plant roots on the erodibility of soils are clearly evident from the data of Weaver (25) in Table 51. His experimental technique for measuring the erodibility

TABLE 51

THE EFFECT OF PLANT ROOTS ON THE ERODIBILITY OF SOILS (WEAVER, 25)

Crop	Dry weight, gm.	In per cent of native grasses	Time of erosion, min.
Big bluestem.....	462	..	60*
Bluegrass.....	282	..	60*
Hungarian brome grass, 4-yr. old...	220	48	120 plus 25 min. with nozzle attached
Alfalfa, 4-yr. old (15 per cent of roots fibrous).....	196	43	21
Sweet clover, 2-yr. old.....	66	18	15
Sudan grass.....	77	17	20
Winter wheat.....	75	16	17
Hegari sorgo.....	74	20	10

* Bare soil eroded in 18 min.; erosion time for bare soil in other experiments was 7 to 10 min.

of soils consisted in subjecting small samples of soils to the eroding action of water that was applied at the rate of 12.7 gallons per minute under a pressure of 1 pound per 1.5 square inches from an open hose. The samples of soils in their natural structure were taken to a depth of about 10 cm. (4 inches); the total surface area was one-half a square meter.

These data point out the differences that may be expected between various crops with respect to their effects on holding the soil together against the eroding action of the rain and runoff water. The sorghum plants were 3.5 feet tall when the sample was taken and were thick in the row. After the tops were removed, however, the soil eroded almost as easily as if it were bare. Most intertilled crops, such as

corn, tobacco, cotton and soybeans, afford little protection for the soil after the tops are removed. The roots of winter wheat extended the erosion time to 17 minutes.

The Sudan grass was ungrazed and was 5 feet high when the soil was sampled. The root system, however, was only about 17 per cent as extensive as native grasses. The erosion time was increased to 20 minutes. Two-year-old sweet clover eroded in about the same time as Sudan grass. Four-year-old alfalfa, with a total root system equivalent to 43 per cent of that under native grasses, extended the erosion time to only 21 minutes. This small resistance to erosion in the presence of such a large root system was probably due to the fact that only about 15 per cent of the total roots were of a fibrous nature.

The interesting grass in this table is the Hungarian brome grass. Although the root system was only 48 per cent as large as the native grasses, the erosion time was more than double that of the bluestem sods. This high resistance to erosion was due to the dense mat of interwoven rhizomes. Such root systems are effective soil binders.

The root effects of plants on soil erosion is excellently summarized by the following statements of Weaver with respect to prairie grasses:

Prairie plants are largely grasses and the parts underground consist of roots, rhizomes, and bases of stems. As a result of the interlacing and clutching of the particles of earth by the myriads of roots of grasses, the soil is compressed into granules and these are surrounded by colloidal films which maintain their permanence and stability. Living underground parts in prairie compose, by weight, approximately one-tenth of the total organic matter. These plant parts slowly deteriorate, die and decay with the passing of the years, only to be replaced by new ones. The channels made by roots in pressing aside the soil remain for a long period of time, greatly increasing the pore space, and consequently the absorption of water. They may readily be traced for many feet, far into the soil and subsoil. Recent studies have shown that a cubic foot of soil has many miles of root fiber. About one-half of the underground plant parts, by weight, occur in the surface 6 inches of soil of prairie. The close dependence of root materials upon abundance and vigor of the tops cannot be overemphasized. Frequent removal of tops weakens the roots and may cause them to die. In prairie, it is only when the vegetation is much trampled and closely grazed and the amount of underground parts greatly diminished that serious erosion begins.

BIOLOGICAL INFLUENCES. It is a well-known fact that the soil fauna are most abundant where there is ample vegetative cover. This is particularly true of dense sods and good forest covers. The soil under a thick forest litter is permeated with the channels of earthworms,

beetles and other biotic life. These channels make the soil rapidly permeable to water. Removal of the litter by burning or pasturing destroys the basic condition that favors the development of an abundant soil flora. The permeability of the soil disappears along with the decreasing biotic activity. This fact is strikingly confirmed by the data of Auten (1), who found that the rate of absorption of water by burned woods was only about 20 per cent as fast as in the normal undisturbed oak woods.

Biotic channels are also present under good grass vegetation but are probably not as important in determining the structure of the upper soil layers as in forest soils.

Wollny (26) was one of the first to demonstrate the structure-improving effects of biotic activity. He studied the changes in soil structure that were effected by earthworm activities. The total porosity was increased by earthworm channels from 57.1 per cent to 59.8 per cent; the non-capillary porosity was increased from 8.9 per cent to 31.2 per cent. Permeability for water was raised from 2.9 liters per 10 hours to 74 liters after earthworms had been introduced into the soil. The channeled soil was about $2\frac{1}{2}$ times more permeable to air.

TRANSPIRATION EFFECTS. After the soil has drained to about its field capacity after a rain, the major removal of water from the soil is through the transpiration of vegetation. When one considers that 7 to 9 inches of water are transpired to produce a ton of hay per acre and 10 to 12 inches to produce a 50-bushel crop of corn or oats, it is immediately evident that vegetation increases the storage capacity of the soil for rainfall. When the transpiration effect on the soil-moisture supply is combined with the interception of rainfall, decreased velocity of runoff and increased soil porosity phases of the vegetation factor, the importance of a protective vegetative cover becomes all the more significant.

There is one point that should be emphasized in regard to the vegetation factor in controlling runoff. That is, a vegetative cover is most effective in decreasing the amount and velocity of runoff when it is growing. During the winter months in the temperate zone, when the large areas of leaf surface are not present to intercept the raindrops and to transpire water, vegetation functions primarily by decreasing the rate of runoff. It may also protect the soil from freezing and thereby maintain a fairly good infiltration rate for the soil in question. In this case, however, water eventually reaches the streams as ground-water flow since the plants are not utilizing water. Therefore, this effect is one of decreasing the velocity of runoff, thought of in terms of the influence of runoff on flood control. In light of these facts, we

should not generalize too much on the relation of the vegetation factor to runoff when this vegetation is not active.

The Soil Factor. The effects of soil properties on water erosion are manifested in two ways. First, there are those properties that determine the rate with which rainfall enters the soil and, second, there are those properties which resist erosion during runoff. Although these two phases of the soil factor are definitely related, the former is by far the more important.

SOIL PROPERTIES AFFECTING INFILTRATION AND RUNOFF. Rainfall that reaches the soil surface (ground rainfall according to Horton (10)) enters the soil, runs off or is detained in small depressions on the surface. Most of the water so detained eventually enters the soil although some of it may be lost by direct evaporation. It is obvious, therefore, that the ability of the soil to absorb rainfall and transmit it through the profile is of great importance in determining the amount of runoff.

The Concept of the Infiltration Capacity of Soils. Horton (10), an internationally recognized authority on hydrology, has proposed and developed the concept of the infiltration capacity of soils as one of the basic steps of analyzing runoff hydrographs. The reader is referred to his publication on the analysis of the hydrograph for an introductory discussion of surface runoff phenomena (11).

Horton (10) has defined **infiltration** as the process involved when water soaks into the ground. In other words, infiltration refers to that process whereby water enters the environment of the soil through the immediate surface. It is not synonymous with percolation. The latter refers to the movement of water through the profile; infiltration refers to the movement of water into the soil. It is true that percolation is generally involved in the process of infiltration but this is not necessarily the case.

The rate at which water can enter the soil is called the **infiltration capacity**. Horton recognizes both a maximum and minimum infiltration capacity of soils. For any given rain, the maximum value obtains at the beginning of the rain. It decreases rapidly at first, owing to changes in the structure of the surface soil, and then gradually reaches a somewhat stable minimum. The maximum infiltration capacity is rarely equivalent to the percolation rate or transmission capacity. The minimum infiltration capacity, however, approaches the percolation rate of the soil profile.

When the rainfall intensity exceeds the infiltration capacity, runoff occurs. Therefore, as the infiltration capacity of a soil decreases from maximum to minimum during a rain, the amount of runoff increases.

Rainfall in intensities below the infiltration capacity is diverted into the soil where it may contribute to the ground-water flow, or return to the air as vapor through evaporation and transpiration or become a part of the capillary-water content of the soil. From the point of view of water conservation and erosion control, therefore, the ideal hydrologic condition within a soil would be one in which the infiltration capacity would equal the rainfall intensity, and the field capacity would be high enough to retain sufficient water for plant growth. It is obvious, however, that the intensity of certain storms is so high as to make the infiltration of all the precipitation into the soil practically impossible.

Common observation has shown that soils vary considerably in the amount of water which is absorbed during rainfall. Experimental observations at the various Federal Soil and Water Conservation Experiment Stations have provided some data to show the magnitude of these variations. For example, data from the Clarinda (Iowa) and Bethany (Missouri) Stations have shown that on plots cropped to continuous corn, with the rows running up and down the slope, the Marshall silt loam absorbed 95.9 per cent of the annual rainfall and the Shelby loam 72.31 per cent. In other words, the ratio of runoff of the Shelby to the Marshall was approximately 7. Musgrave (18) measured the infiltration capacity of these soils and found that the ratio of the Marshall to the Shelby was 7.8 for a one-hour period. The relationship of the infiltration rate to rainfall intensity for these two soils is shown in Figure 69. These results indicate that runoff on the Marshall could be prevented by any treatment which would impound about 1.25 inches of water. The low infiltration capacity of the Shelby makes it impossible to prevent runoff by any practical means of impoundage. These data clearly illustrate some differences that can be expected in the infiltration capacities of various soils.

The infiltration capacity of a given soil is a very dynamic property. It changes with season and management, although Horton has observed that the maximum and minimum infiltration capacities have fairly definite values at a given season. The more important factors affecting the infiltration capacity are (1) the permeability of the profile, (2) the condition of the soil surface and (3) the soil-moisture content. These are extremely dynamic factors, especially the last two.

Permeability of the Profile and Infiltration. In Chapters V and VI, the importance of non-capillary porosity in determining the permeability of the soil profile was particularly emphasized. For example, Figure 53 in Chapter VI shows that the percolation rate, which is approximately the same as the minimum infiltration rate in the experi-

ments cited, increases directly with the non-capillary porosity. Moreover, if the results in Figure 69 are studied in light of the porosity relationships shown in Figure 36, Chapter V, one is immediately impressed with the importance of non-capillary porosity in determining the permeability of these two profiles. The structure of these two profiles was discussed in Chapter V, page 163. The Marshall silt loam has about 5 times the volume of non-capillary pores as the Shelby

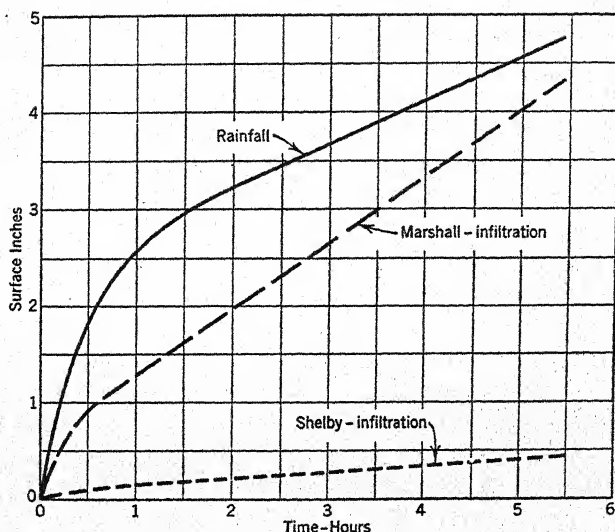


FIG. 69. The infiltration capacities of Marshall silt loam and Shelby loam (Musgrave, 18).

loam. Musgrave found that the infiltration capacity of the Marshall was about 7 times that of the Shelby.

The amount of water percolating through a soil profile is determined by the permeability of the least pervious horizon. If this horizon happens to be in the subsoil, the permeability of the entire profile depends upon the rate of movement through the subsoil. This means that the maximum infiltration capacity of such soils depends upon the ability of the horizons above the subsoil to take in water. Once they become saturated, the percolation rate of the subsoil regulates further water absorption and the minimum infiltration capacity is reached.

If the least pervious layer of the soil is at the immediate surface, the rate of water movement into and through the profile is limited by the infiltration rate of the surface. In this case, the entire hydrology of the profile is associated with water movement through the immediate surface.

Condition of the Soil Surface and Infiltration. The soil structural relationships of the immediate surface have not been sufficiently recognized as key factors in determining the infiltration capacity of soils. This is especially true with respect to the maximum infiltration capacity. This fact has already been discussed in Chapter IX in connection with soil tilth. The data of Neal, in Table 42, and of Musgrave and Free, in Figure 66, clearly show that an open, permeable surface increases the infiltration capacity of the soil. A compact surface is responsible for low infiltration rates and high runoff.

A detailed analysis of the results in Figure 66 shows that cultivating this soil to a depth of 4 and 6 inches increased the maximum infiltration capacity of the soil. This increased infiltration rate, however, was almost completely manifested during the first 15 to 30 minutes after the application of water. After this time interval the rate of infiltration was constant for the uncultivated as well as the cultivated area. In other words, the increased porosity within the upper 4 to 6 inches made possible a rapid absorption of rainfall. When this pore space was filled with water, the infiltration capacity of the soil was determined by the rate of percolation through the lower horizons. This value was approximately the same for the cultivated and uncultivated areas.

The condition of the surface varies with crop, climate and tillage practices. The importance of the vegetation factor has been discussed in previous paragraphs. The greater the vegetative canopy and the more extensive the root system, the greater will be the beneficial effects of plants in maintaining an open structure in the surface. Thus, grasses are more effective than small grains and the latter are more effective than intertilled crops.

Climatic influences are manifested either in the form of reduced or increased infiltration capacities. Torrential rainfall distribution at the time when the soil is not adequately protected causes rapid compaction of the surface and clogging of the pores. The infiltration rate is quickly reduced under the beating action of the rain; runoff is greatly increased. Freezing of the soil surface as well as the upper soil layers reduces the rate of infiltration of winter rains. Consequently, a high percentage of the rainfall runs off. This is one of the factors that contributes to the severity of winter floods.

Prolonged dry spells generally increase the infiltration capacity of soils because of the severe cracking and checking of the soil surface. These cracks, which are caused by soil shrinkage, are often numerous and of considerable size. Rainfall enters the soil very rapidly and runoff is decreased until the cracks are closed as a result of swelling.

It is difficult to evaluate the significance of these cracks in the infiltration of rainfall; however, they are extremely significant with the heavier types of soils that normally possess low infiltration capacities.

Tillage practices that involve the loosening of the soil increase the infiltration capacity. Those practices that have a compaction action tend to decrease the infiltration rate. Soils that are freshly plowed or cultivated absorb water at a greater rate than the same soil previous to the tillage operation. Soils that are rolled or dragged have lower infiltration capacities after the tillage operation. These tillage effects were discussed in more detail in Chapter IX.

Soil Moisture Content and Infiltration. In the discussion of the climatic factor in runoff and erosion problems, attention was called to the significance of the moisture content of the soil, at the time of the storm, on the amount of runoff (3). Wollny (4), in studying the various factors that influence runoff and erosion, recognized the soil moisture as an important factor in determining runoff from slopes with different exposures. He noted that runoff was greater according to the following series: north > west > east > south. The moisture content of the soil increased in the same order.

Horton (10) divided storms into two classes, A and B, upon the basis of their occurrence in relation to other rains. This classification was designed to take into consideration the moisture content of the soil at the time of the rain. Class A storms were those that followed two or more rainless days; the class B storms were designated as those that occurred after one or less rainless days. The infiltration capacity of a loessial soil in Iowa for these two types of storms for the different months is given in Table 52. These results, which represent the infiltration over a complete watershed, show that the rate of infiltration of the rainfall that falls on moist soil was considerably slower than on dry soils. This fact is obvious from a comparison of the infiltration rates for the months of May, June, July and August. It should also be noted that the infiltration for all storms was low during the winter and early spring months. The infiltration rates for the summer months were also increased by tillage operations, as shown by the values for the class A storms during this period. The lower infiltration rates for the class B storms indicate that heavy rainfall compacted the soil and increased the saturation of the pores in the surface horizon with water.

Neal (21) observed that the soil-moisture content at the beginning of a rain had a greater effect upon the rate of infiltration during the first 20 minutes than any other factor. The rate of infiltration varied approximately inversely as the square root of the soil-moisture content

at the beginning of the rain. Diseker and Yoder (6) also found that saturating the soil with water greatly increased the runoff. These results, together with the findings of other investigators, emphasize the need of evaluating the soil-moisture status of the soil profile in any investigation of runoff.

TABLE 52

ANALYSIS OF INFILTRATION CAPACITY OF LOESS SOIL ON RALSTON CREEK AT
IOWA CITY, IOWA

(8-yr. average. Data of Horton, 10)

	Type of storm					
	Class A storms		Class B storms		All storms	
	No. storms	Infiltration, in. per hr.	No. storms	Infiltration, in. per hr.	No. storms	Infiltration, in. per hr.
Feb.	2	0.17	2	0.17
Mar.	3	0.12	1	0.03	4	0.10
Apr.	1	0.12	1	0.10	2	0.11
May	3	1.30	2	0.24	5	0.87
June	9	1.57	8	0.81	17	1.21
July	4	1.36	3	0.32	7	0.91
Aug.	5	0.75	3	0.47	8	0.65
Sept.	6	1.11	1	1.92	7	1.23
Oct.	2	0.25	2	0.25
Nov.	2	0.11	2	0.11

By way of summary, it should be re-emphasized that, excepting conditions of adequate grass or forest protection, the structural properties of the immediate surface and the moisture content of the soil profile, especially in the upper layers, are the two factors that play the major roles in determining the amount of runoff from any given storm. Inability to evaluate these factors properly have hindered the progress of flood-control research. Recent developments in techniques for measuring the soil moisture and structure promise to increase our knowledge of runoff factors in erosion and flood control.

Methods for Measuring Infiltration. Infiltration investigations have been undertaken by numerous workers. Several different types of techniques have been proposed to measure the infiltration capacity of a soil. These techniques may be divided into two main categories: (1) considering the infiltration capacity to be the difference between

rainfall and runoff rates and (2) considering infiltration to be the rate at which water will enter the soil under experimental conditions where runoff is prevented.

The first concept of using the difference between rainfall and runoff as an index of the infiltration capacity has been developed by Horton (10, 11). After taking into consideration that part of the water which is detained on the ground, after the period of rainfall excess has ended, and enters the soil as infiltration, he has been able to calculate the infiltration capacity of an entire watershed from rainfall and runoff data.

Pearse and Bertelson (23) and others have attempted to modify this technique by using small plots instead of large watersheds. The plots of Pearse and his associates were only one square foot. Water was supplied by a small overhead system to simulate natural rainfall. Various members of the Soil Conservation Service have used artificial rainfall on plots as large as $\frac{1}{100}$ acres. The major difficulty of the small plots is the lateral movement of water away from the plot border as infiltration proceeds. This movement tends to give high infiltration capacities. The major difficulty of the large plots is the size of apparatus that provides the artificial rain.

The second type of measurements of the infiltration capacity has at least three variations. They are the use of (1) long tubes that are sunk into the soil to prevent lateral movement of water during infiltration, (2) shallow cylinders or metal boxes that permit free lateral movement of water as it is applied to the soil and (3) buffer areas to compensate for the lateral movement of infiltrated water.

The technique of Musgrave (18) is a good example of the metal cylinder for measuring infiltration. A metal cylinder is jacked into the soil to the desired depth and water is supplied from a burette in such a way as to maintain a constant head of water on the surface. The rate of infiltration is obtained from the burette readings. This technique is rather simple and gives a fairly close index of the minimum infiltration capacity after the soil column within the tube has become saturated. It does not allow the ready escape of entrapped air which tends to decrease the infiltration rate. It does not cause an appreciable disturbance of the soil surface during the addition of water which tends to maintain a higher infiltration capacity than under natural conditions. Neal (21) observed that the cylinder technique gave values for the infiltration capacities of the Putnam silt loam from 2 to 3 times higher than those obtained as the difference between rainfall and runoff. These differences were due in part to the decrease in the infiltration rate with the application of a 3-inch rain on the surface. The structure of the surface was changed under the impact of the raindrops.

The second variation in determining the infiltration capacity by the rate at which water is absorbed under conditions of free lateral flow is illustrated by the technique of Auten (1). Four successive one-liter portions of water are poured into a frame one foot square that is sealed to the soil surface with wet clay. The time of infiltration for each portion is recorded. A variation in this technique consists of forcing the rectangular frame into the soil to a depth of one or two inches. Since water is applied to a relatively small area that is dry on all sides, considerable lateral flow takes place. The use of successive amounts of water aims at obviating this difficulty. There is every opportunity for entrapped air to escape. There is little disturbance of the surface during the application of the water.

The third variation within this second type of measurement aims at compensating for the factor of lateral flow by making use of buffer compartments in the infiltration apparatus. The technique of Kohnke (12) illustrates this type of measurement. A sixteen-compartment rectangular infiltration apparatus is inserted into the soil to a depth of one to two inches. Each compartment has an area of 200 sq. cm. Water is added to each compartment simultaneously from each of sixteen 500-cc. burettes. The flow of water is regulated so that only about one-third of the soil area is submerged; the other two-thirds are occasionally sprinkled. This technique permits the ready escape of air. Readings are taken every 5 minutes but only the rate of infiltration of the four inner compartments is used as representing the true infiltration capacity of the soil. The purpose of the outer compartments is to wet the soil so that the lateral flow of water from the inner compartments will be restricted. Although this technique overcomes some of the objections of the tube method, there is reason to believe that the method gives results that are much higher than the infiltration that occurs under natural conditions.

Infiltration techniques are as yet in their experimental stages. There is as yet no particular method that does not have some important disadvantage. Artificial-rain applications on small areas offer considerable promise, if the effect of lateral flow can be overcome.

Soil Properties Affecting Erosion during Runoff. The major soil properties that reduce the amount of erosion per unit of runoff are those that increase the soil's resistance to dispersion. There is little concrete information on this particular phase of the erosion problem, with the possible exception of root effects. Common observation has shown that granular soils erode. For example, the well-granulated Marshall silt loam in northwestern Missouri has suffered serious erosion. Continuous cropping of corn, with the rows running up and

down the slope, has been the primary cause for the severity of the erosion. Nevertheless, there are certain storms that will make the most ideal soil erode, if it is not protected with vegetation.

Miller and Krusekopf (17) have reported that corn following corn caused almost twice as much erosion per unit runoff as corn following clover. No soil structure data are available to explain the exact cause of these differences. It is safe to conclude, however, that the soil was more permeable after the clover crop as evidenced by much less total runoff. The fact that the density of runoff was decreased by the plowing under of clover suggests either a greater stability of granules or a binding together of the soil particles by the clover roots. Musgrave and Norton (20) have shown that the addition of manure to fallow land reduced the density of runoff. The presence of the organic matter increased the soil's resistance to dispersion and movement.

Diseker and Yoder (6) analyzed the character of the eroded material from Cecil clay and found that aggregates rather than textural separates were removed in the erosion process. The diameters of these aggregates were often larger than those of coarse sand.

The investigations of Diseker and Yoder and of Neal indicate that there are two structural conditions of the soil surface that determine the amount of erosion. When the soil is dry and somewhat compact, the first increment of rain causes a slaking action at the immediate surface and a high density of runoff. As the rain continues, after this thin layer of loose soil has been removed, a wet compact surface is produced which decreases the density of runoff in spite of a greater total runoff. The resistance of the wet layer apparently increases with the clay content.

On the other hand, if the soil is loose and granular, such as in freshly cultivated fields, there is little erosion at the beginning of the storm. The infiltration capacity is high and runoff and soil losses are low. However, when the rainfall intensity exceeds the infiltration capacity, the soil losses are high. There is no binding together of the granules and the soil erodes to the bottom of the loosened layer. Granulation in this case expedites erosion instead of hindering it.

Therefore, it seems that at the beginning of large, intense storms, or throughout short storms, soil is eroded as a result of the slaking and beating action of raindrops on the soil. Fine material is carried away during this phase of the erosion process. The more granular and resistant the soil to slaking and dispersion, the lower will be the density of runoff. As the storm continues, however, and runoff increases, both in amount and velocity, the erosion of the soil depends upon the coherence of the particles in the immediate surface with those

underneath. In this case, a highly granular soil will probably erode more per given amount of runoff than one that has a smooth, compacted surface.

The greater erodibility of loose aggregates as compared with a compacted surface is well illustrated in exposed clay subsoils. When such exposed areas are wet there are rather extreme compaction and coherence between particles; the density of runoff is not high. However, let a prolonged dry spell appear with a few occasional showers and the surface of these areas will slake down into a coarsely aggregated condition. The next heavy rain will easily remove this loosened layer down to the compacted zone.

In addition to these types of surface conditions, there is the situation of extreme wetness that is often encountered with highly silty soils on rather moderate slopes. During periods of continued rainfall with low intensities, these soils become saturated with water; cohesion is very small. If a torrential rain should fall on such a soil, runoff would be high and erosion would be extremely severe. This condition exists often with the rolling claypan soils of Missouri and adjacent states.

Thus, it is seen that aggregation, volume changes and hydration play important parts in affecting erosion once runoff occurs. The experimental data clearly point out that vegetation is the only reliable factor to cause significant resistance to erosion during runoff. The fact that the most granulated soils severely erode under intense storms that cause high runoffs suggests that soil properties are more important in minimizing runoff than in lessening erosion after runoff begins.

EROSION AS A FUNCTION OF SOIL PROPERTIES. Soil scientists have been particularly interested in obtaining an index of the erodibility of soils by measuring some physical properties or group of properties of the soil. The investigations of Middleton (16) were among the first attempts in this direction. He measured a variety of physical properties of the soils from the various Federal Erosion Experiment Stations and searched for some correlations between these properties and the erosion that was determined in the field. He suggested the dispersion ratio as an index of the ease with which the particles would be brought into suspension by the action of the rain or by runoff water. This ratio was obtained by dividing the amount of silt plus clay that was easily suspended by shaking the soil in pure water by the total quantity of silt plus clay that was present. The greater this ratio, the more easily the soil could be dispersed.

The colloid-moisture equivalent ratio was used to express the relative permeability of the soil for water. Permeability was considered

to increase with this ratio. On the theoretical assumption that erosion should increase directly with the dispersion ratio and inversely with the colloid-moisture equivalent ratio, the erosion ratio was obtained. The erosion ratio was simply equal to dispersion ratio/colloid-moisture equivalent ratio. Middleton and his associates found rather satisfactory qualitative correlations between the erosion ratio and the erodibility of most of the soils that were investigated.

Lutz (15) made a detailed study of the physical and chemical properties of the Davidson clay and the Iredell sandy clay loam in order to find out the causes for the differences in erosion of these two soils. Field observations have definitely pointed out that the Iredell is an erosive and the Davidson a non-erosive soil. The two soils occur in adjacent areas; all the factors that generally affect erosion, excepting soil properties, are practically the same on the two sites. The B-horizon of the Davidson as it occurs in the field always appears friable, granular, porous and permeable to water, whereas the B-layer of the Iredell is distinctly plastic, sticky and impervious. Inasmuch as the clay content of the Davidson is higher than the Iredell, the differences in erosiveness cannot be explained on the basis of texture; the mechanical composition of the two soils should favor higher erosion in the Davidson.

Laboratory studies of the physical properties of these soils showed conclusively that one of the principal differences between the Davidson and Iredell from the standpoint of soil erosion is the extent to which the soils are aggregated. The Davidson contains a large amount of aggregates practically all of which are larger than 0.25 mm. in diameter; the Iredell contains a somewhat lower quantity, most of which are relatively small compared with those of the Davidson. Microscopic examinations have shown that the Davidson aggregates are composed of clusters of smaller granules and therefore are permeable, whereas the Iredell aggregates are dense and impermeable. The greater permeability of the Davidson profile is the direct result of the higher content of large aggregates, which are more friable, porous and stable than those of the Iredell. The greater stability of the Davidson aggregates is indicated by the extent to which the soil is naturally granulated and by the resistance of the soil to dispersion as evidenced by a lower dispersion ratio.

Studies with the extracted colloid showed that the Davidson is flocculated when saturated with H, Ba, Ca, Li, Na or K ions; additions of NaOH up to 150 per cent of the saturation capacity failed to produce dispersion. On the other hand, only the H-, Ba-, and Ca- Iredell colloids were flocculated. The state of flocculation of the colloids is evidence of the high state of aggregation of the natural Davidson soil.

Swelling data indicated that the Davidson colloidal material does not swell, irrespective of the nature of the exchangeable cations present. Colloids extracted from the Iredell are highly hydrated and undergo considerable swelling.

The results help to explain the differences between these two soils. The erosiveness of the Iredell must be due chiefly to the ease with which it is dispersed and its impermeability to water. The non-erosive nature of the Davidson is due to the high state of flocculation of its finer mechanical separates into large, porous and stable granules. These granules resist dispersion and permit a rapid percolation of water through the soil profile. The physical properties which affect permeability and the ease of dispersion are the paramount factors determining the erodibility of soils.

The effect of soil factors on erosion may be summarized in the following descriptive equation:

$$E = K \frac{D}{APp} \quad (52)$$

where K refers to a proportionality constant, involving the other factors mentioned in equation 51, D is an index of the ease of dispersion, A is an expression of the infiltration capacity of the soil surface, P characterizes the permeability of the soil profile and p denotes the size of the soil particles. The equation emphasizes the need for evaluating both the dispersity of the soil particles in the surface layers and the pore-space relationships of all layers of the profile before an approximate picture of the erodibility of a soil can be had.

THE IMPORTANCE OF SOIL PROPERTIES IN THE ENGINEERING ASPECTS OF EROSION CONTROL

The engineering aspects of erosion control are primarily concerned with changing slope characteristics so that the amount and velocity of runoff are lessened. For example, the length of slope may be controlled by the spacing of terraces across the slope. The effect of degree of slope on the concentration of runoff water may likewise be controlled by terracing. The cutting action of concentrated water in terrace channels, gullies or drainage ways may be controlled by a combination of suitable engineering structures and vegetative protection. There are certain soil properties that play an important part in these engineering practices. They will be mentioned briefly.

Soil Properties and Terracing. The major effect of terraces on the control of erosion is to cut down the length of the watershed and

to conduct the water off the field at a low velocity. Reducing the velocity of the water in the terrace channel causes a deposition of a large portion of the silt load in the channel. Consequently, only a small part of the soil that is eroded between terraces leaves the field in the water that caused the erosion. When terracing is coupled with good agronomic practices, erosion can be decreased to a minimum.

Terraces are not as effective in controlling the total runoff as they are in slowing up the rate of runoff. Hydrograph charts from an

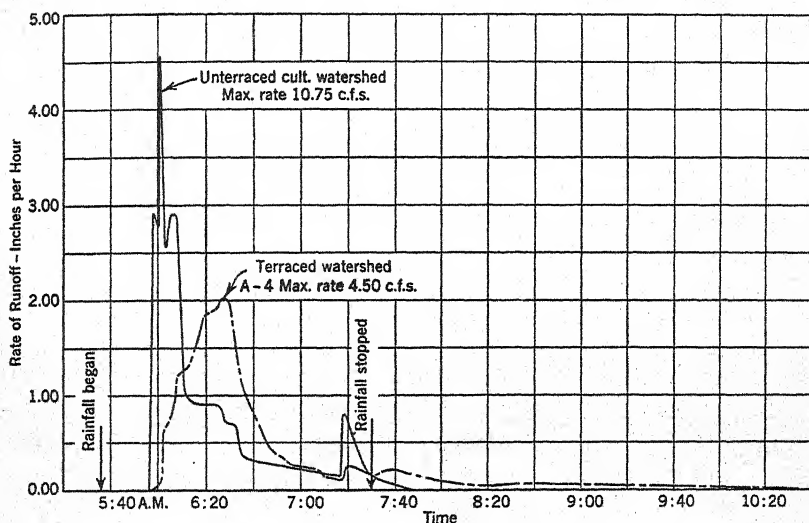


FIG. 70. Hydrograph curves showing runoff from one storm at LaCrosse, Wisconsin (Hays and Palmer, 8).

unterraced cultivated watershed and from a variable grade terrace, for the rain of July 5, 1935, at LaCrosse, Wisconsin, clearly illustrate this point. (See Figure 70.) The total runoff from the unterraced watershed was 68.5 per cent of the rainfall; this runoff produced 51.16 tons of soil loss per acre. Runoff from the variable grade terrace was 66.3 per cent of the rainfall; 7.8 tons of soil were lost per acre. What were the reasons for such large differences in soil loss when the total runoff was nearly the same? In the first place, it is granted that the amount of soil deposited in the silt boxes at the end of the terrace did not represent the total amount of soil that was moved during the rain. Nevertheless, the soil that was deposited in the terrace channel was not lost from the field. The answer to this difference is found in comparing the rates of runoff from the two areas. Rapid runoff occurred on the unterraced area about 18 minutes after the rain began;

rather slow runoff occurred on the terraced area about 7 minutes later. The maximum rate of runoff on the untterraced land was 10.75 cubic feet per second; that on the terraced land was 4.5 cubic feet per second. Runoff from the untterraced area stopped about 18 minutes after cessation of the rain; runoff from the terrace did not stop completely until about 3 hours after the rain. These differences show that the rates of runoff are extremely important in the dispersion and transportation of soil.

Significant reductions in the total runoff apparently occur only on permeable soils. This fact is shown in Table 53. These results indi-

TABLE 53

EFFECT OF TERRACING ON RUNOFF FROM VARIOUS SOILS

(Data from progress reports of Federal Soil and Water Conservation Experiment Stations)

Place	Slope in per cent	Rainfall in inches	Relative permeability of soil	Percentage runoff	
				Terraced	Untterraced
Guthrie, Okla.....	5.1	31.3	slow	18.6	22.5
Tyler, Texas.....	7.5	45.7	slow	16.5	14.0
LaCrosse, Wis.....	13.1	26.8	fair	4.9	8.9
Bethany, Mo.....	6.7	31.7	very slow	18.7	19.2
Pullman, Wash.....	20.0	26.6	rapid	11.2	31.0

cate that the total runoff from terraced land is not significantly different from that on untterraced areas in the case of the less permeable soils. Runoff is markedly decreased by terracing on permeable soils because the reduced velocity of runoff allows more time for infiltration.

TERRACE SPACING. Terrace spacing is determined primarily by the slope of the land. Tables have been set up as a basis for establishing the correct vertical spacing for a certain slope. The data in Table 53 as well as results reported by Musgrave and Norton (20) indicate that terrace spacing should vary with the ability of the soil to absorb water and transmit it through the profile. If the infiltration capacity of a soil is twice that of another soil under comparable slope conditions, a wider spacing of terraces on the former than on the latter should not cause increased runoff losses. This fact is strikingly illustrated by the data from the Soil and Water Conservation Experiment Station

on the Marshall silt loam (20). The infiltration capacity of this soil is so high that runoff from terraces with a vertical spacing of 6 feet on a 9 per cent slope did not average as much as 0.2 per cent of the total rainfall during the first 5 years of experimentation. The usual recommendation of this type of slope is a vertical spacing of 4 to 5 feet. It is possible that the normal recommendations for an extremely impervious soil may often prove too high because of the low infiltration capacity of such soils. It is encouraging to note that the aforementioned soil differences are being considered at present to a greater extent in terrace-spacing recommendations than they have been in the past.

LEVEL TERRACES. Recommendation of level terraces has been a rather standard practice for the conservation of rainfall. This recommendation has considerable merit in areas of limited rainfall on soils that are readily permeable. The practice becomes somewhat dangerous on soils that have a low permeability. In these impermeable soils, water concentrates in the terrace channel and is not carried away by infiltration into the soil and percolation downward through the profile. Any vegetation that may have been planted in the terrace channel is thereby destroyed because of inadequate drainage. Consequently, the permeability of subsurface layers is a very important factor that will determine the success or failure of level terraces.

TERRACE OUTLETS. Every agricultural engineer emphasizes to the farmer that the terrace outlet is the key to the successful functioning of a system of terraces. Terrace outlets must handle concentrated water. Therefore, they must be completely stabilized. The ease and economy of stabilization depend considerably upon the properties of the soil over which this concentrated water must flow. The major difficulties in obtaining suitable outlets are found in connection with the terracing of moderately sloping soils that have a high silt content and on soils that possess heavy clay subsoils.

Many silty soils do not contain sufficient clay to cause much cohesion between particles. Such soils erode easily under the cutting action of concentrated water. They present a difficult problem in the construction and maintenance of good terrace outlets. Since the soil has little cohesion, binding power can be provided through the roots of suitable grasses.

A different problem exists with claypan soils. Unfavorable fertility and physical conditions make the establishment of vegetation difficult. Even when good vegetative protection has been provided, a prolonged drought may kill a considerable portion of the vegetation. The stability of the outlet is immediately reduced. The occurrence of cracks during dry weather also weakens the stability of the outlet.

The suggestions offered in Chapter V for the physical improvement of clay soils have been found to aid materially in the production of good vegetation on such soils.

Soil Properties and Contour Farming. Contour farming is essentially a system of small terraces around the slope. Each row of a particular crop is a miniature terrace. Therefore, the success of contour farming for causing increased infiltration of water should be greater on permeable soils, even as terraces are more successful on these types of soil. The results in Table 54 indicate that this relationship holds. The Marshall silt loam at Clarinda, Iowa, was able to absorb all the rainfall when the corn was planted in lister furrows on the contour. Running the furrows up and down the slope permitted as much runoff on this permeable soil as on the other two less permeable soils. Although the differences in total runoff are not as great in the other soils, the amount of erosion was markedly reduced by contouring.

TABLE 54

RUNOFF FROM CULTIVATED CROPS PLANTED ON THE CONTOUR
(Data from progress reports of Federal Soil and Water Conservation Experiment Stations)

Place	Slope in per cent	Rainfall in inches	Relative permeability of soil	Percentage runoff	
				Rows up and down the slope	Rows on contour
Spur, Texas.....	0.5	18.5	slow	8.8	6.8
Guthrie, Okla.....	6.8	35	slow	13.7	11.3
Clarinda, Ia.....	8.4	26	rapid	11.5	0

The discussion in this chapter calls attention to the complexity of the erosion problem. A thorough analysis and knowledge of all contributing factors are essential to a complete understanding of the problem. All the factors are closely interrelated. Each can be broken down into its various aspects and analyzed in a manner similar to that followed for the soil factor.

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